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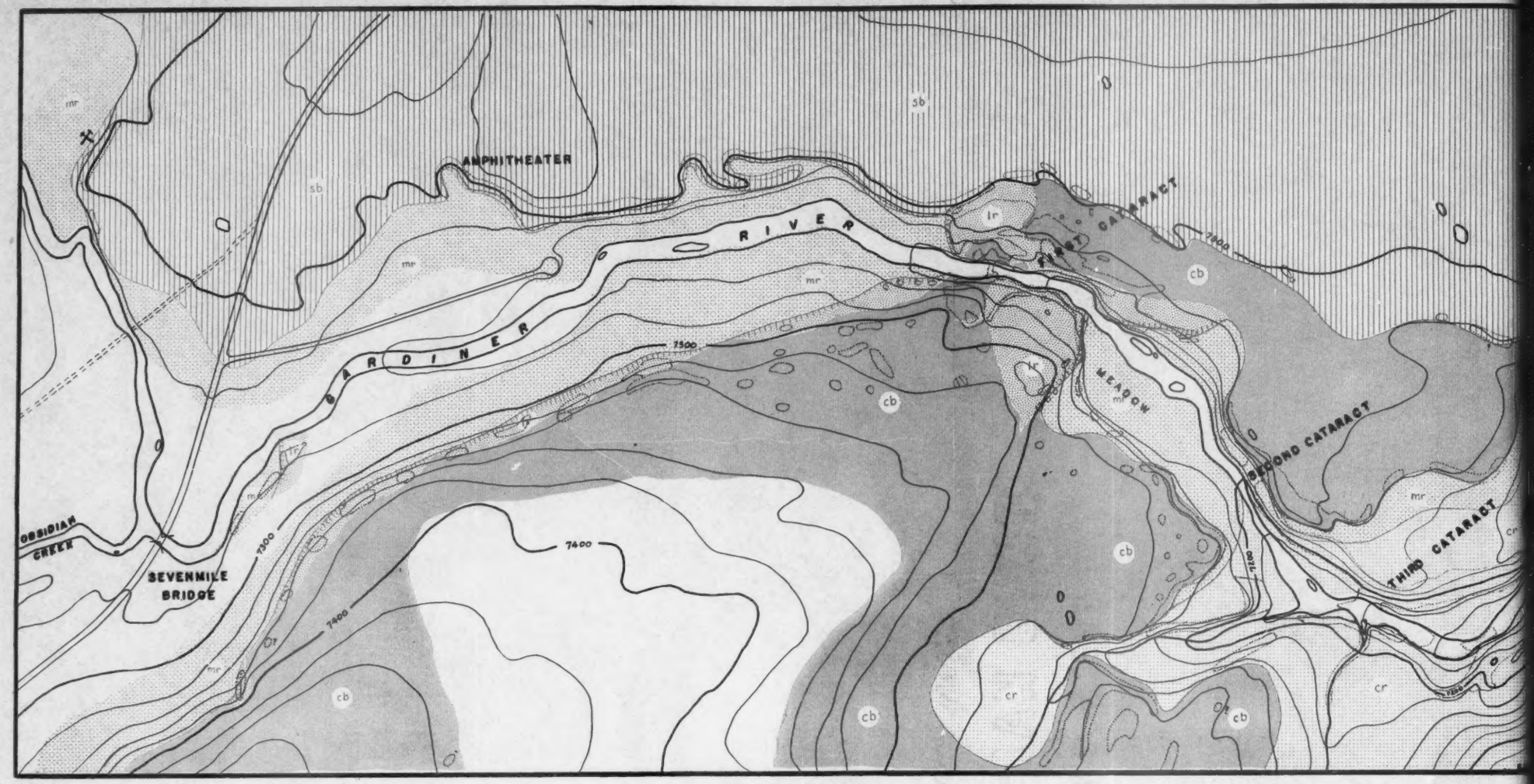
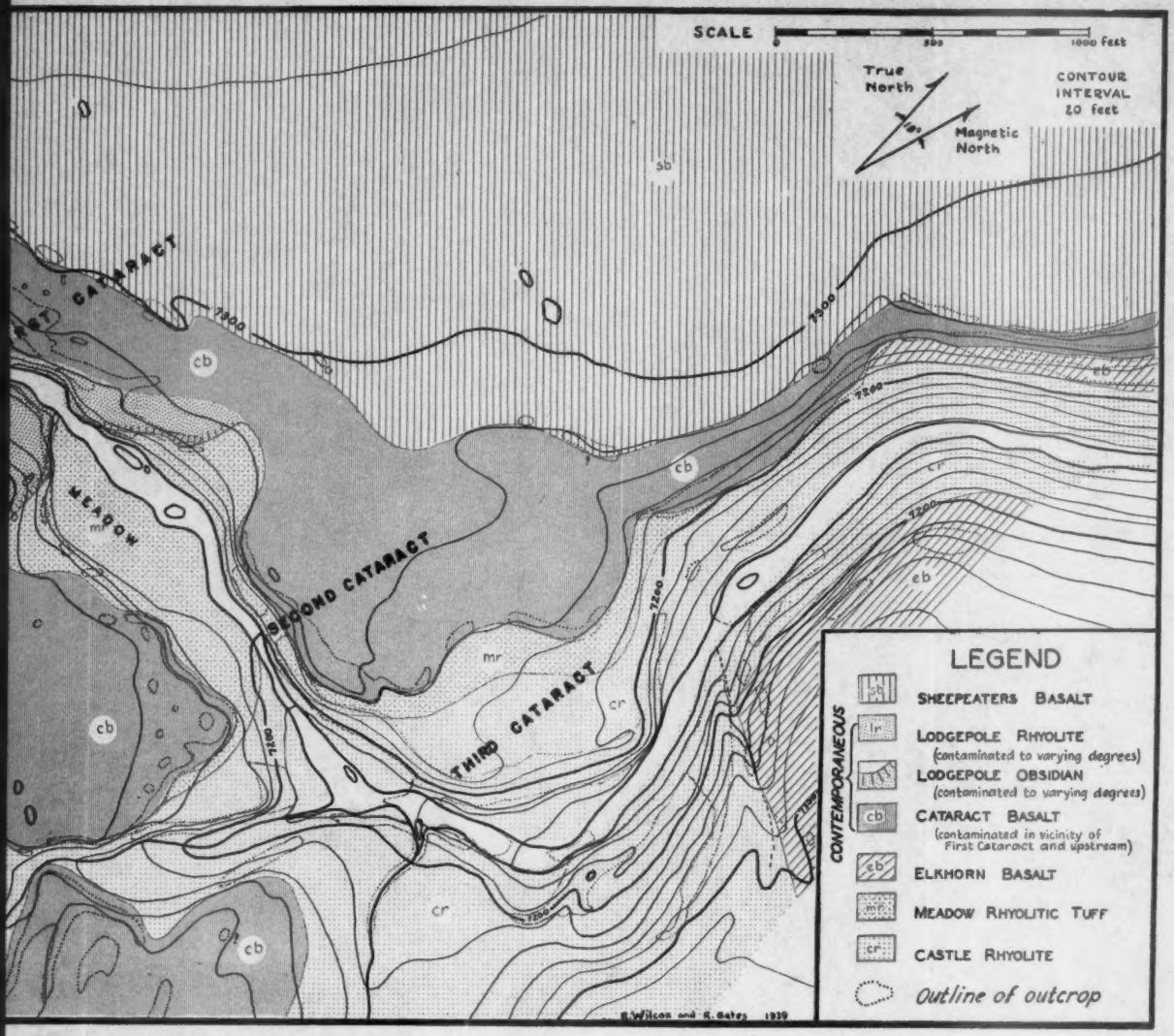


GEOLOGY AND TOPOGRAPHY OF SEVENMILE BRIDGE REGION

Gardiner River, Yellowstone Park, Wyoming



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RHYOLITE-BASALT COMPLEX ON GARDINER RIVER, YELLOWSTONE PARK, WYOMING

BY RAY E. WILCOX

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ABSTRACT

Certain lavas on Gardiner River in Yellowstone Park, Wyoming, have been cited (Iddings, 1899, p. 430; Fenner, 1934; 1937; 1938) as representing extensive assimilation of solid basaltic material by molten rhyolitic lava at pressures near that of the atmosphere. The writer, however, concludes that the features of the Gardiner River lavas resulted from the mixing of nearly contemporaneous rhyolitic and basaltic lavas. On this basis, the occurrence does not bear on the question of the relative importance of assimilation versus gravitational differentiation in the origin of rocks.

The proponents of assimilation base their case on the definite intrusive relationships of the rhyolite into the basalt (Iddings, 1899, p. 430–32; Fenner, 1938), the roughly V-shaped form of the rhyolite-basalt contact, and the chemical gradation from rhyolite to basalt (Fenner, 1938). However, the writer concludes that, if the two lavas had been erupted simultaneously, the solidifying effect of the normally cooler rhyolitic lava on the basaltic lava would have caused the same relationships.

Two important features point strongly to an origin of the Gardiner River lavas by mixing of molten rhyolitic and basaltic lavas: (1) A chill phase of the rhyolite against country rock but not against the basalt indicates that the basalt was not cold when contacted by the rhyolitic lava. (2) The presence of quartz and orthoclase xenocrysts (originally phenocrysts of the rhyolitic magma) within the basalt several yards from the contact indicates that the basalt was liquid or mushy when contacted by the rhyolitic lava.

INTRODUCTION

In his description of the igneous geology of Yellowstone Park, Iddings drew attention to a mass of complexly intermingled rhyolite, basalt, and rock of intermediate composition on Gardiner River. The origin of this mass he attributed to the assimilative action of a rhyolite lava on an older basalt (Iddings, 1899, p. 430–432). Subsequent literature on assimilation cites this locality repeatedly as a forceful argument for the ability of siliceous magmas to produce far-reaching changes in adjacent rocks, even after the magma has been erupted on the surface and its pressure reduced nearly to that of the atmosphere. Fenner (1934; 1937; 1938), after a more complete examination of the Gardiner River complex, applied to it the same mode of origin as Iddings but emphasized the role of rhyolitic emanations in the assimilation of the basalt.

The present investigation began in 1938 with a brief field examination and petrographic study in which were revealed several features that did not appear to support assimilation. Because of the critical nature of this occurrence, it was decided to map the area, examine it in detail, and attempt to study the evidence objectively. The writer concludes that the data do not support the claim that molten rhyolitic lava has assimilated cold basalt. Rather, it is concluded that the basalt and rhyolite were erupted simultaneously and mingled while both were still hot.

This conclusion, if true, rules out highly energized rhyolitic emanations as the effective agents in assimilation of the basalt. It means that the Gardiner River occurrence does not furnish valid evidence of the ability of rhyolitic lavas (or salic lavas

in general) to assimilate country rock and cannot justifiably be introduced into the long-standing controversy over the relative importance of assimilation versus gravitational differentiation in the origin of rocks. The writer wishes to make clear, however, that the conclusions in this paper are not put forward as arguments against the ability of granitic magmas at depth to accomplish significant assimilation of country rock.

Field work on the Gardiner River area was done during 7 weeks in 1938 and 1939. In addition, 11 weeks in the summer of 1940 were spent in that vicinity on an extension of the same problem, the results of which are to be published later.

ACKNOWLEDGMENTS

Field expenses in the 1940 season were borne in part by a grant from the Penrose Fund of The Geological Society of America and by the Yellowstone-Beartooth Research Association. The writer is indebted for field assistance during the 1939 season to Mr. R. M. Gates and in the 1940 season to Messrs. R. M. Gates, D. S. Turner, R. C. Townsend, A. R. Perko, W. Ross, C. Ingersoll, and T. Baker. Much aid and numerous courtesies were rendered during the course of the field work by members of the U. S. Forestry Service and of the U. S. National Park Service, especially by Dr. C. Max Bauer, Chief Naturalist of Yellowstone Park.

Petrographic study was done in the laboratories of the Department of Geology of the University of Wisconsin during 1938, 1939, and 1940. Valuable critical discussion was given by the students and staff of the Department of Geology of the University of Wisconsin, especially by Prof. R. C. Emmons and Prof. S. A. Tyler, and much aid in the preparation of the manuscript was given by the writer's wife. The Department of Geology of Northwestern University kindly furnished facilities for the drafting of maps and diagrams during the summer of 1943.

FIELD RELATIONS

GEOLOGIC SETTING

The Gardiner River rhyolite-basalt complex lies in the northern part of a Tertiary volcanic field extending across northwestern Wyoming into Idaho. Iddings (1896, p. 614) pointed out that the general order of extrusion was (1) the andesites of the Absaroka Mountains, (2) the rhyolites of Yellowstone Plateau, and (3) the basalts of the Snake River Plains. Although the major portions of the basaltic extrusions followed cessation of the rhyolitic extrusions, there was an intermediate period during which the two materials were being erupted alternately and, in some localities, at extremely short intervals.

The Gardiner River rhyolite-basalt complex is part of the lava filling of a valley eroded in the main mass of the Yellowstone Plateau rhyolites. Prior to the formation of the rhyolite-basalt complex a shallow valley extended north-south across the site of the complex, transverse to the present gorge. After the filling of this valley by the mingled rhyolitic and basaltic lavas and after a period of erosion, another basalt spread over the area from the northeast, lapping against the higher slopes of the complex and earlier tuffs and rhyolites south of the present Gardiner River (Pl. 1).

The edge of this latest flow was especially vulnerable to erosion, and along it was cut much of the present course of the Gardiner River in the area under consideration. Along some stretches the river had reached the crumbly earlier tuffs and rhyolites but could not cut deeper because of the dam of massive rhyolite-basalt lying across the river's course at First Cataract. At this point the river slowly cut the narrow gorge where today the mingled rhyolite and basalt are so strikingly exposed.

RHYOLITE-BASALT COMPLEX

The main body of the complex lies 5 miles south of Mammoth Hot Springs and three-fourths of a mile downstream from Sevenmile Bridge (Fig. 1; Pl. 1), at a point where the Gardiner River begins to descend its canyon on the north edge of the plateau. At this point the river has exposed a nearly endless variety of structures and textures, combining rhyolitic and basaltic material, for about 900 feet along the walls and upper slopes of First Cataract Gorge.

The structures and rock types in the gorge have been described in detail by Fenner (1938), and the writer would add emphasis only to certain features which appear to be critical in determining the origin of the complex. For details of the complex the reader is referred to the map of the First Cataract area in the present paper (Pl. 2) and to the comprehensive description by Fenner (1938, p. 1449–1468).

The term Lodgepole rhyolite has been assigned to the rhyolite flow that contributed to the formation of the complex. This flow may have been extensive, but as yet no large masses uncontaminated by basaltic material have been found in the region. Lithologically, the least contaminated Lodgepole rhyolite is a normal gray to buff rock of lithoidal texture carrying abundant phenocrysts of orthoclase and quartz, grading at the extreme east and west ends of the complex into glassy and spherulitic rock (Fenner, 1938, p. 1449, Pl. 2, fig. 3). A narrow zone of outcrops of contaminated rhyolite and obsidian extends southeasterly up the Gardiner River for about two-thirds of a mile (Pl. 1).

The basalt of the complex is here referred to as the Cataract basalt. Its normal facies is a massive, medium-grained basalt, carrying abundant phenocrysts of plagioclase and ferromagnesian minerals. It occurs in its uncontaminated phase at Second Cataract, about 700 feet downstream from the complex at First Cataract, and outcrops are found at frequent intervals several miles southward up the grade of a wide tributary valley. In contrast to the younger Sheepeaters basalt, to be described later, the Cataract basalt is almost lacking in columnar structure (Pl. 3, fig. 1).

The variations in the structures of the rhyolite-basalt combinations in the complex at First Cataract fall into three general classes:

(1) Xenolithic structures are common and include both rounded and angular fragments of basalt in all stages of mechanical and chemical disintegration. These are involved in a groundmass of rhyolite which may itself be changed toward a more basic composition (Pl. 3, fig. 3; also Fenner, 1938, Pl. 1, figs. 2, 3; Pl. 2, fig. 2; Pls. 7, 8). Clear-cut examples of xenoliths of rhyolite in a basalt matrix have not been found.

(2) Rhyolitic dikes, ranging from a few millimeters up to several feet in width, intrude the basalt (Pl. 2; also Fenner, 1938, Pl. 4, fig. 3; Pl. 5, fig. 2). Some contacts of

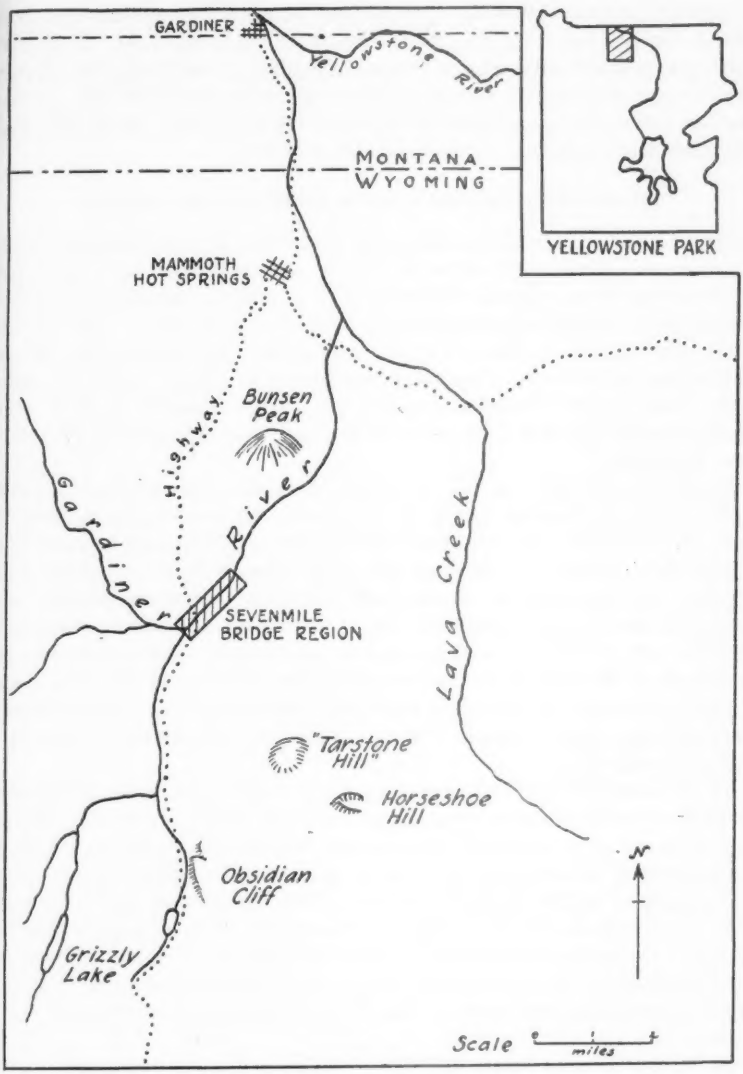


FIGURE 1.—*Location of rhyolite-basalt complex*
Sevenmile Bridge region

the dikes are sharp and straight; others, especially the narrower dikelets, are transitional and irregular. Clear-cut dikes of basalt cutting rhyolite have not been found.

(3) "Marble cake" and "zebra" structures, in which the rhyolitic and basaltic

portions are drawn out in streaks or twisted into whorls of light and dark shades, are present over portions of the complex (Fenner, 1938, Pl. 8, figs. 4, 6). Basaltic material appears as isolated swirls and clots in the rhyolitic groundmass, while all gradations between this structure and true xenolithic structures are observable. In many cases the rock in rough hand specimen may appear to be a nearly homogeneous basalt, but, when polished, light and dark streaks are revealed.

FORMATIONS ASSOCIATED WITH THE RHYOLITE-BASALT COMPLEX

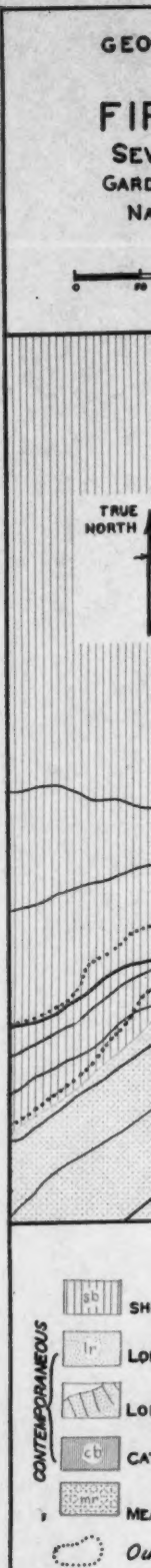
General statement.—Detailed mapping both upstream and downstream from the occurrence was carried out to determine the age relations of the rocks of the complex to the country rocks. Toward both east and west ends of First Cataract gorge the texture of the rhyolitic portions changes from lithoidal to glassy. The marginal positions of these glassy phases of the complex immediately suggest chilled selvages such as develop normally at the margins of extrusive or shallow intrusive igneous rock. Since a glassy texture might, however, indicate operation of other factors, careful consideration will be given to the relationship of this marginal area to the older formations.

Meadow rhyolitic tuff.—As the east end of the First Cataract is approached the walls of the gorge flare out sharply to encompass a flat marshy meadow about 400 feet wide (Pl. 2; Pl. 4, fig. 2). Forty feet beyond the easternmost exposure of the rhyolite-basalt complex on the north side of the meadow is an outcrop of crumbly rhyolitic tuff, called here the Meadow tuff. Sixty feet beyond the easternmost exposture of the complex on the south side of the meadow is another outcrop of the Meadow tuff. Additional outcrops occur along the steep banks still farther east, and all along the banks of the meadow this tuffaceous material can readily be uncovered by gouging out humus and roots with a hammer or pick. Downstream the walls converge again at Second Cataract where the tuff extends underneath the Cataract basalt flow.

The Meadow tuff is older than the rocks of the complex, as indicated by the physiographic and spatial relations, even though the actual contact is covered. The position of the contact on the north side of the river is limited to a span of 40 feet between the outcrops of contaminated obsidian on the west and rhyolitic tuff on the east. The outcrop of obsidian is higher, and flow planes within it dip gently toward the west, away from the tuff, indicating that the obsidian lies on a gently inclined tuff surface. The sudden flaring of the walls of the gorge at its eastern end and large slumped blocks of obsidian indicate extensive undermining by removal of an easily eroded material (*i.e.*, the crumbly Meadow tuff) immediately underneath the obsidian.

At Second Cataract, where the walls of the gorge are uncontaminated Cataract basalt (the basaltic end member of the complex), exposures are better, and the contact between the Cataract basalt and the underlying Meadow tuff can be exposed by scraping the soil away.

These relationships indicate that the Lodgepole rhyolite, obsidian, and Cataract basalt of the complex are younger than the Meadow rhyolitic tuff and that the obsidian phase of the rhyolite-basalt complex is the chilled border, lying against slightly inclined surfaces of the older tuff.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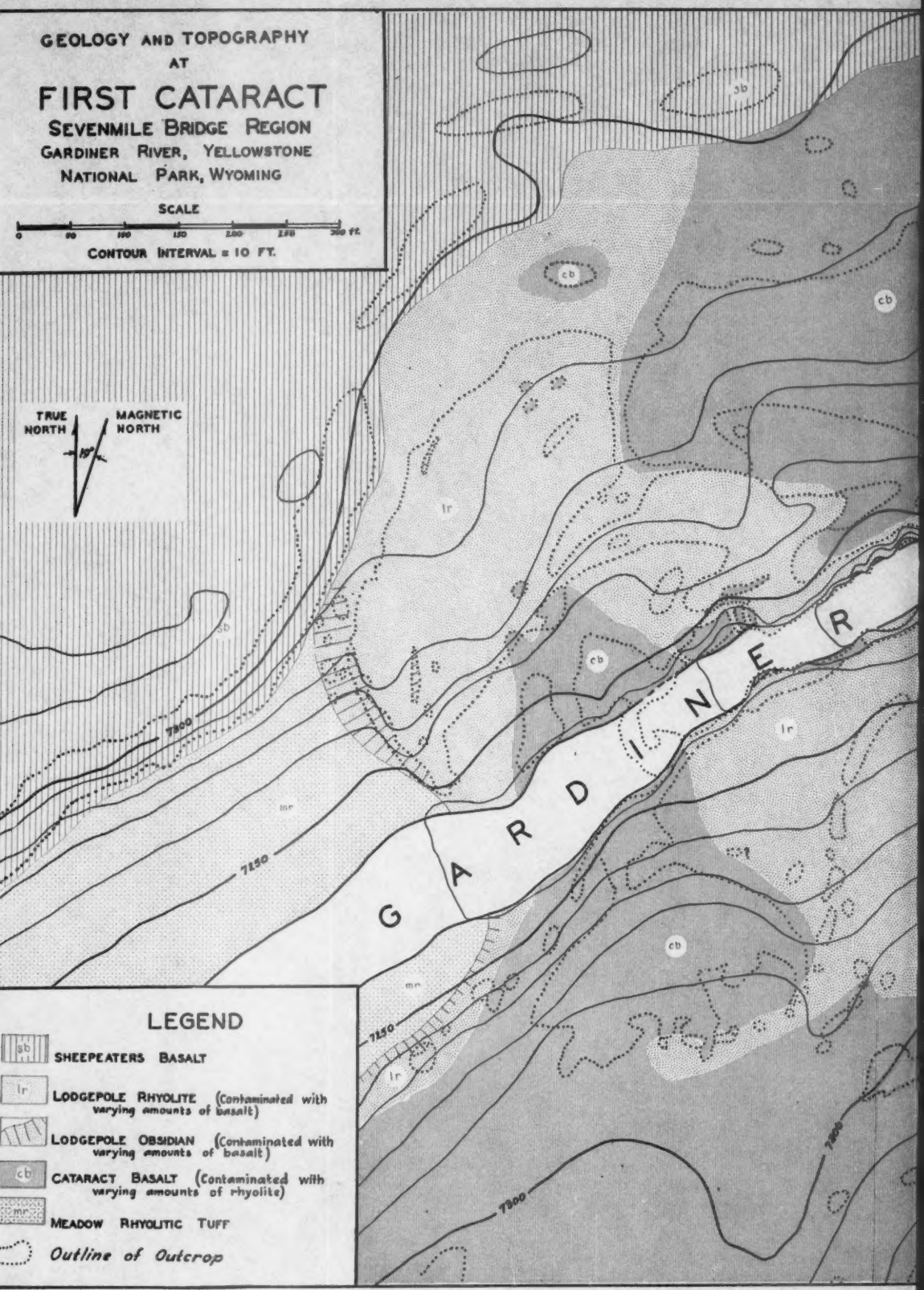
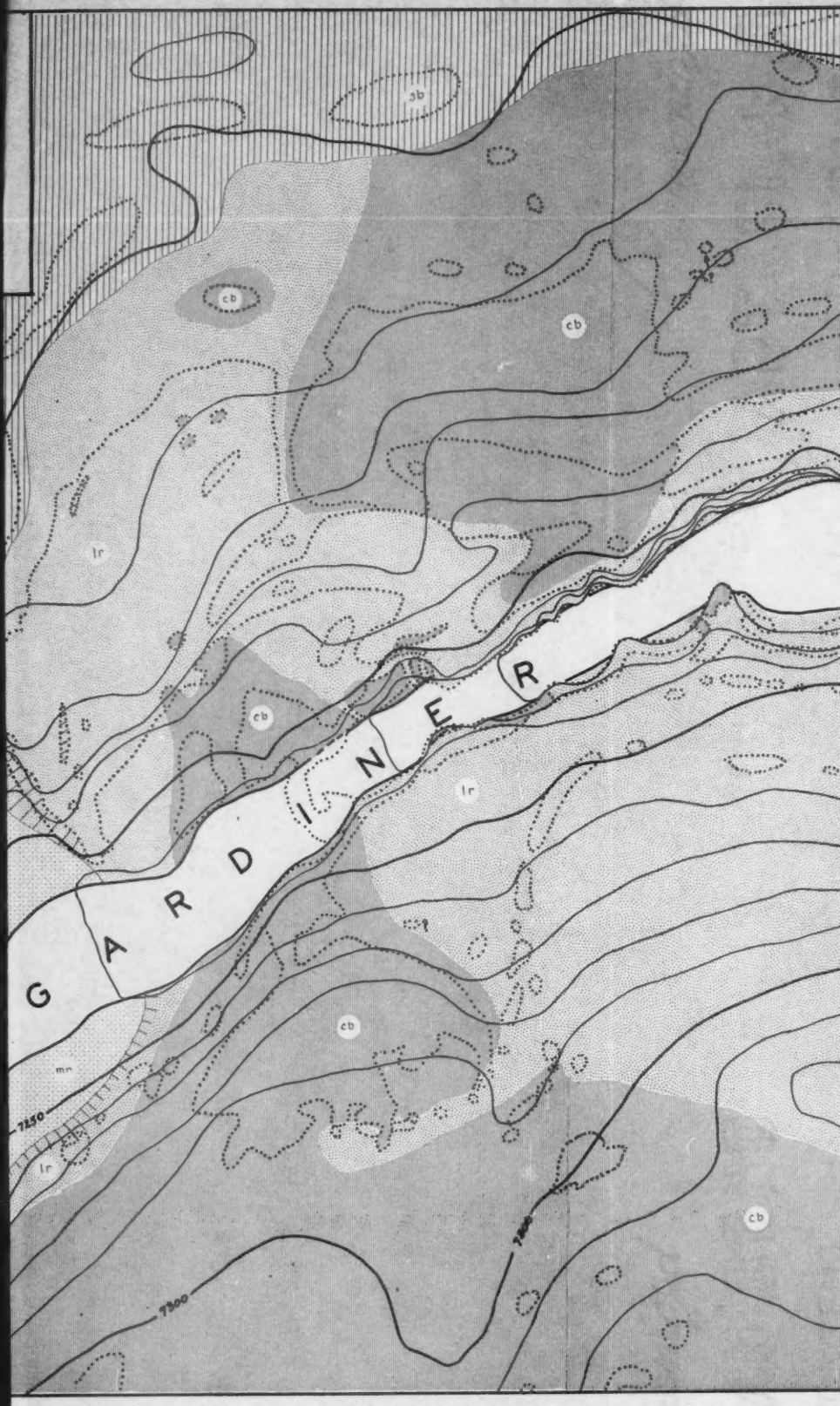
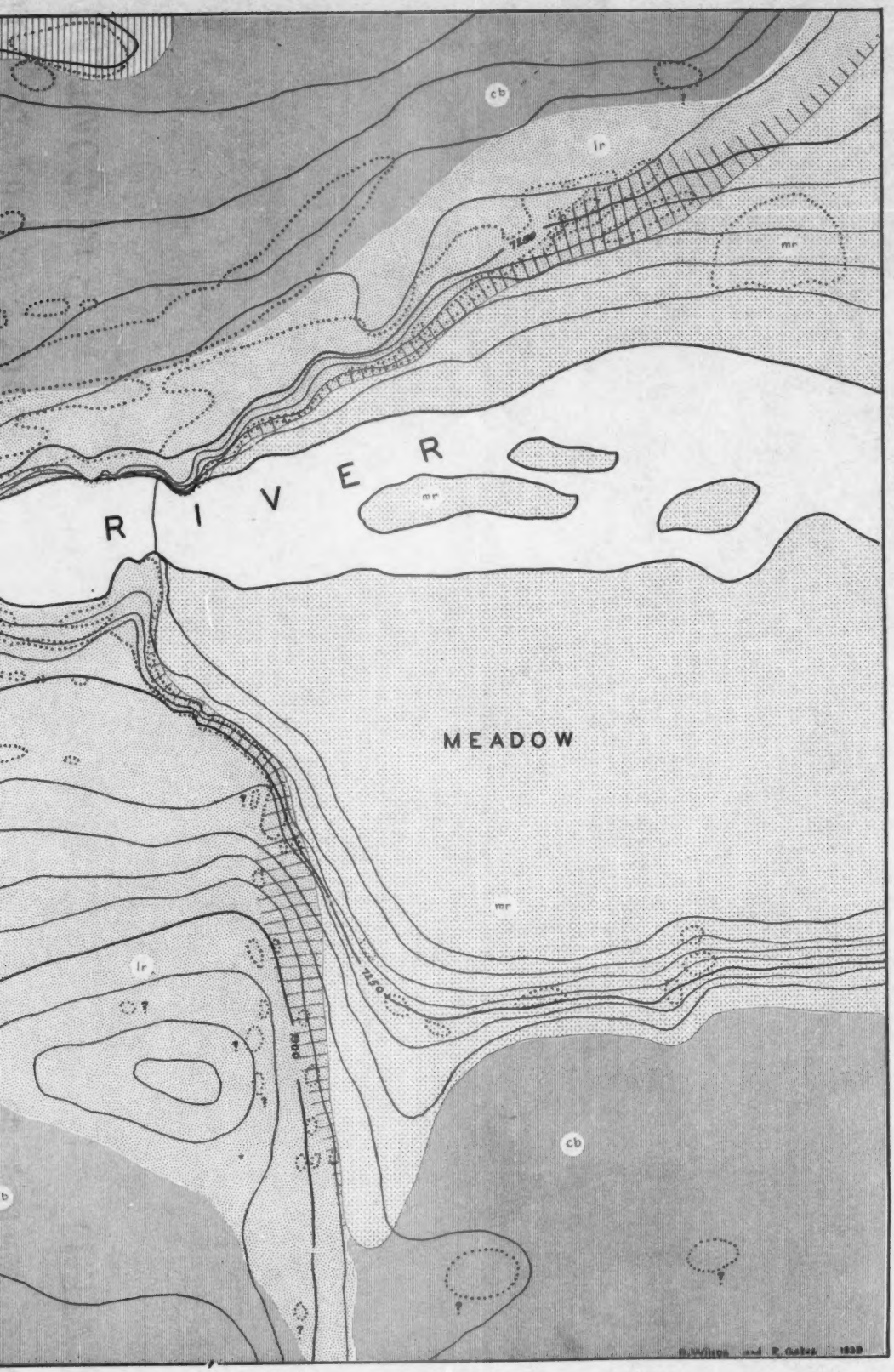


GEOLOGY AND TOPOGRAPHY
Sevenmile Bridge region, Y



GEOLOGY AND TOPOGRAPHY OF FIRST CATARACT

Sevenmile Bridge region, Yellowstone Park, Wyoming



TARACT AREA
, Wyoming

Castle rhyolite.—Below Third Cataract a compact rhyolite, the Castle rhyolite, underlies the Meadow tuff, the latter being traceable to this point. The dense layered Castle rhyolite (Pl. 4, fig. 1), exposed in cliffs 100 feet high at the point below Third Cataract where the Gardiner River turns northward (Pl. 4), is like much of the rhyolite of the western half of Yellowstone Park. Although underlying the Meadow tuff, it may have been an earlier part of the same eruptive stage as the Meadow tuff. The layering of the Castle rhyolite, accentuated by weathering, is apparently due to differing degrees of coherency and porosity rather than to differences in composition. The Castle rhyolite is megascopically and microscopically similar to the so-called "welded tuffs" of other regions (Marshall, 1935; Grange, 1937; Gilbert, 1938), but the origin of the Castle rhyolite and the related rhyolites in other parts of Yellowstone plateau is not clear. Certain features, the discussion of which is outside the scope of this paper, lead to the tentative conclusion that the Yellowstone formations were emplaced as true flows rather than as hot ash falls that might have been compacted and welded.

Sheepeaters basalt.—Besides the basalt in the rhyolite-basalt complex (the Cataract basalt), there are two additional basalt flows in the vicinity of First Cataract. Fenner (1938, p. 1445), considering the basalt flows of the region to have been emplaced and deeply eroded before the formation of the complex, makes no distinction among the separate flows. Detailed field mapping and petrographic study have shown, however, that one basalt flow is older than, and another younger than, the basalt of the complex.

The younger flow will here be called the Sheepeaters basalt because of its typical exposures along the Sheepeaters Cliffs on the north side of Gardiner River. It has an even-grained (nonporphyritic) texture and striking columnar structure (Pl. 3, fig. 2; Fenner, 1938, Pl. 1, fig. 1). It blankets the entire northwest portion of the mapped area (Pl. 1), outcropping quite consistently at elevations near 7300 feet at the top of the cliffs along the northern side of the river. This basalt forms the walls of "The Amphitheater," an extinct waterfall of a glacial tributary of Gardiner River just below Sevenmile Bridge. Its outcrop continues along the left bank almost without interruption to First Cataract where it apparently overlaps a portion of the complex.

Downstream from First Cataract the outcrops of Sheepeaters basalt are widely spaced, but the continuation of the edge of the formation can be traced northeasterly by means of a low but persistent shelf about 5 feet in height. This shelf meets the rim of the canyon about half a mile below First Cataract, and the flow again outcrops at intervals for almost another mile downstream, still showing the perfect columnar structure and even-grained texture.

The Sheepeaters basalt, along the northern edge of the rhyolite-basalt complex at First Cataract (Pl. 2), is the "rim basalt," spoken of by Fenner as the unmodified phase of the basalt that participated in the complex and considered by him to be the same flow as that which outcrops at Second Cataract, 1000 feet downstream (Fenner, 1938, p. 1447, 1450). Both the "rim basalt" and the basalt of Second Cataract can be traced, however, to the cliff at the extreme northeastern end of the mapped area (Pl. 1), where the "rim basalt" overlies the basalt of Second Cataract (the Cataract basalt). This field evidence and petrographic evidence to be discussed in a later

section indicate that the Sheepeaters basalt is younger than the Cataract basalt and did not participate in the formation of the rhyolite-basalt complex.

Elkhorn basalt.—Underneath the Cataract basalt at the extreme northeastern end of the mapped area (Sevenmile Bridge region, Pl. 1) is exposed the oldest of the three basalts, a porphyritic rock, slightly reddish in some phases and locally carrying numerous amygdules of zeolite. It is referred to here as the Elkhorn basalt. Although its lower part is quite dense, it grades upward into scoriaceous and cindery basalt near the contact with the overlying Cataract basalt.

The material underlying the Elkhorn basalt is not revealed, but the abrupt decrease in steepness of the canyon wall just below the outcrop may indicate a more easily eroded material beneath, perhaps the Meadow rhyolitic tuff or the Castle rhyolite. Since not far upstream and downstream from this point the Castle rhyolite is exposed below the basalts, it has been so mapped. On the basis of the relationships described, the Elkhorn basalt would therefore be intermediate in age between the Castle rhyolite and the Cataract basalt.

PETROGRAPHY

RHYOLITE-BASALT COMPLEX

General relations.—The uncontaminated Lodgepole rhyolite of the complex at First Cataract carries abundant phenocrysts, chiefly of quartz and orthoclase, in a groundmass that is variously granophyric, lithoidal, spherulitic, or glassy. The uncontaminated Cataract basalt carries abundant phenocrysts of bytownite, olivine, and pyroxene in a groundmass of plagioclase, pyroxene, and subordinate magnetite.

The intermediate mixtures of rhyolite and basalt are as complex in thin section as in the field. Basaltic masses, many seemingly on the point of disintegration, are scattered through a rock that is mainly rhyolitic. Large xenocrysts of bytownite, olivine, and pyroxene likewise occur abundantly in the rhyolitic portion, many of them entirely isolated from basaltic material. Very few islands of rhyolitic material in basaltic matrix are seen in thin sections, although large quartz and orthoclase grains are common.

Lodgepole rhyolite (rhyolitic end member).—It is difficult to obtain a specimen of rhyolite known to be free from contamination by basaltic material. The description, therefore, is of thin sections of rhyolite taken from outcrops containing relatively few xenoliths of basaltic material and in which little alteration is noticeable.

The most common groundmass texture in the rhyolite is a fine-grained intergrowth of quartz and feldspar with a micrographic intergrowth visible in some of the coarser phases. Phenocrysts of quartz and orthoclase attain 1.2 mm. in diameter. A very few phenocrysts of sodic plagioclase are present, and most carry a shell of orthoclase. Nearly all phenocrysts show rounded corners and slight embayment, as is common in rhyolites.

The glassy phase of the rhyolite carries abundant microlitic crystals and spherulites in addition to quartz and orthoclase phenocrysts. The phenocrysts in the glass are identical with those of the normal crystalline rhyolite, except that the rounding and embayment is, in some cases, less marked. As in the normal phase of the rhyolite, a

FIGURE 1. CATARACT BASALT
At Second Cataract.

FIGURE 2. SHEEPEATERS BASALT
Above First Cataract.

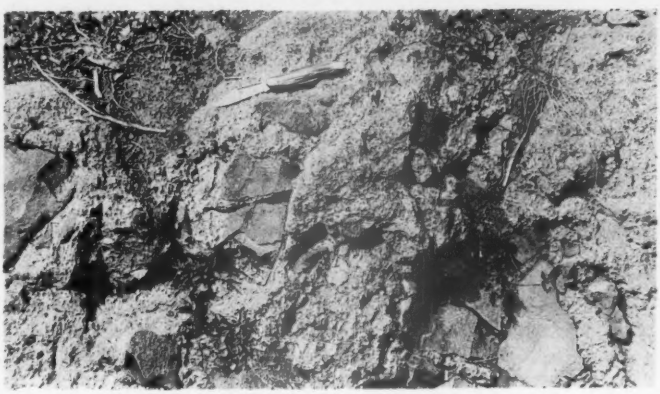


FIGURE 3. BASALTIC XEONLITHS IN RHYOLITE
Note incipient disintegration of some xenoliths.

OUTCROPS ON GARDINER RIVER

FIGURE 1. CASTLE RHYOLITE
Below Third Cataract.

FIGURE 2. MEADOW
Looking downstream toward Second Cataract from First Cataract. Outcrop of rhyolite-basalt complex in left foreground.

OUTCROPS ON GARDINER RIVER

few sodic plagioclase phenocrysts, commonly with a rim of orthoclase, are present in the obsidian (Pl. 5, fig. 1). The twinning bands of some of the feldspars have the patchy and imperfect development that characterizes soda orthoclase (Pl. 5, fig. 2). It seems evident that these nuclei of plagioclase and soda orthoclase were obtained, probably as xenocrysts from some external source, by the rhyolitic magma long enough before the extrusion of the magma to permit corrosion and subsequent addition of orthoclase.

Cataract basalt (*basaltic end member*).—The groundmass of the uncontaminated Cataract basalt is composed of interlocking plagioclase laths up to 0.2 mm. in length associated with pyroxene and magnetite granules up to 0.03 mm. in diameter. Phenocrysts include bytownite, olivine, pyroxene, and some magnetite (Pl. 7, figs. 1, 2; Pl. 8, fig. 2). The presence of these abundant phenocrysts in the Cataract basalt contrasts with the nearly total absence of phenocrysts in the Sheepeaters basalt (Pl. 8, fig. 4).

The bytownite phenocrysts are as much as 0.9 mm. in length and 0.3 mm. in width. In most specimens they possess nearly perfect crystallographic terminations, but in some cases, notably in the exposure on the north side of Second Cataract, the ends of the bytownite phenocrysts are frayed and show other evidences of slight resorption (Pl. 7, fig. 2). Euhedral and anhedral olivine phenocrysts of the Cataract basalt attain a diameter of 0.2 mm. Most olivine phenocrysts show typical alteration on their margins and along fractures.

The groundmass plagioclase of the uncontaminated Cataract basalt, determined by extinction angles, ranges from $Ab_{35}An_{65}$ to $Ab_{30}An_{70}$, with the albite content generally greater on the margins than in the interior. The plagioclase phenocrysts, determined by orientation on the universal stage and by refractive indices of cleavage fragments, are bytownite and range from $Ab_{30}An_{70}$ to $Ab_{26}An_{74}$. Indistinct oscillatory zoning, present in many crystals, shows a general increase in albite content toward the margins, amounting in a few extreme cases to as much as 10 per cent.

Some bytownite phenocrysts of the basalt contain rounded nuclei in which the extinction angle differs radically from the outer portions of the crystals (Pl. 7, fig. 3). These nuclei are much more albitic than the surrounding plagioclase of the crystal, some having a composition of $Ab_{57}An_{43}$, as determined on the universal stage.

Olivine phenocrysts are high in magnesia. Direct measurements on three olivine phenocrysts picked at random gave $2V(-) = 85°–87°$, indicating a composition of $For_{78}Fay_{22}$. Crystals in two other samples revealed $2V(-) = 82°$, a composition of $For_{70}Fay_{30}$. Refractive indices of two large grains of olivine from a glomeroporphyritic group were (for the D line of the spectrum):

Grain No. 1:	Grain No. 2:
$\alpha = 1.6711$	α, not determined
β, not determined	$\beta = 1.6898$
γ, not determined	$\gamma = 1.7064$

These indices indicate a composition of $For_{83}Fay_{17}$ (Winchell, 1931).

Ferromagnesian minerals in the groundmass are chiefly pyroxene. Measurements of optic angles gave $2V(+) = 41°–44°$.

On the basis of field relations, the "rim basalt" has already been designated as a flow separate from the Cataract basalt flow of the complex, and its separate identity is

further established by petrographic characters. The rim basalt was regarded by Fenner, however, as an unaltered phase of the basalt that took part in the formation of the complex, and as part of the flow exposed at Second Cataract (Fenner, 1938, p. 1450). In discussing the mode of alteration of the basalt by rhyolite, Fenner says (1938, p. 1449):

> "The unaltered basalt is a dark-gray, even-grained brittle-looking rock, in which almost no phenocrysts are discernible. Metamorphism by contact with the rhyolite has affected the basalt to various degrees. The first effect noticeable is that the rock appears spotted with numerous small, light-colored phenocrysts. A study of thin sections shows that (with slight exceptions to be noted) new phenocrysts (or porphyroblasts) do not seem actually to have been developed, but most of the groundmass has broken down into a fine-grained, and hence darker, aggregate, leaving some of the original crystals relatively intact and more conspicuous."

A comparison of the relative grain sizes and textures in the basalts from the two masses shows, however, that, while the plagioclase grains of the rim basalt attain a length of only about 0.3 mm., the plagioclase phenocrysts of the contaminated basalt attain a size of 0.9 mm., and, while the plagioclase grains of the rim basalt are anhedral, the plagioclase phenocrysts in the contaminated basalt are all nearly perfectly euhedral. Similar sharp contrasts exist in the sizes and shapes of the ferromagnesian constituents of the two rocks. It does not seem that the strikingly porphyritic texture of the contaminated basalt could have been produced by a simple breaking down of certain portions of the even-textured rim basalt as suggested by Fenner (1938, p. 1449) for the constituents of the rim basalt are neither large enough nor sufficiently perfect in shape to correspond to the phenocrysts of the contaminated basalt. Rather, these features support the field evidence that the rim basalt is an entirely separate flow from the basalt involved with the rhyolite of the complex.

Rocks of intermediate composition in the complex.—The progressively greater contamination of the basalt by rhyolite is seen under the microscope in the increasing albite content of the groundmass plagioclase. Phenocrysts of plagioclase show no change in composition, nor any change in their other characteristics except where they are isolated in rocks predominantly rhyolitic in composition, and even there the only effect on the plagioclase "xenocrysts" is a more intense fracturing and in some cases a rounding of the corners of the crystals. The progressively greater contamination of rhyolite by basalt first shows itself in the appearance of clots and specks of dark material as well as isolated xenocrysts of bytownite, olivine, and pyroxene. Resorption of quartz and orthoclase phenocrysts, with concomitant formation of their reaction products, becomes marked upon advanced contamination, and, where isolated in rock of predominantly basaltic composition, these xenocrysts appear only as "ghost" relics.

To describe completely the petrography of the manifold combinations of rhyolite and basalt that form the complex would extend this paper excessively. Since their variations are continuous, however, a description of several stages in the series gives a general picture. Representatives will be described (1) in order of increasing contamination of the Cataract basalt by rhyolitic material, and (2) in order of increasing contamination of the Lodgepole rhyolite by basaltic material. It is of critical importance in the following to keep clearly in mind the difference between the effects of contamination on the groundmass constituents and on the phenocrysts of the basalt.

Contamination effects first become apparent in the groundmass of the basalt. Extinction angles of the groundmass plagioclase of the slightly contaminated basalt indicate a greater albite content than those of the uncontaminated basalt, and this is generally accompanied by a decrease in grain size of the groundmass. Determination of groundmass ferromagnesian constituents in these and in the greatly contaminated basalts is impractical because there are no readily ascertainable crystallographic directions for extinction angles such as there are in the plagioclase series. In addition their small size and many inclusions make refractive-index determinations in immersion liquids unreliable. If a sufficiently clear grain should be found, there is no assurance that it might not be a fragment of a phenocryst. Plagioclase phenocrysts are, however, of the same composition, size, and shape as those of the uncontaminated basalts, as shown by orientation on the universal stage. A few rounded grains of quartz, up to 0.5 mm. in diameter, present in the basalt, characteristically are surrounded by a corona of fine-grained augite and glass. There are a few wrinkled and turbid remnants of orthoclase crystals.

With increase in rhyolitic material the groundmass plagioclase becomes more albitic, and the grains decrease further in size, while the plagioclase phenocrysts maintain their characteristic composition, size, and euhedral form. Olivine phenocrysts commonly show a thin rim of alteration to a fine-grained, highly birefringent mineral. A few scattered irregular patches and streaks of micrographic quartz and orthoclase blend into the more basic groundmass. Associated with these, as well as isolated in the basic groundmass, are large rounded grains of quartz and orthoclase, more common here than in the only slightly contaminated basalts. The quartz grains are again surrounded by coronas of augite, and many of the orthoclase grains, clear in their centers, are turbid in their outer portions.

The textures of the rocks predominantly rhyolitic in composition are characterized, in general, by the presence of small basaltic xenoliths as well as whorls that show all stages of disintegration and blending into the rhyolite. The groundmass of the small basaltic xenoliths is generally so fine-grained and poorly crystallized that the identification of its feldspars by extinction angles is not feasible. In many cases the groundmass of the basaltic xenoliths is rendered nearly opaque by the abundance of very finely divided iron oxide (magnetite?). Some patches of basaltic material fray out at the edges, and many fragments apparently were just on the point of engulfment by the rhyolite when final solidification occurred. In others the basaltic material has completely disintegrated, and small dustlike particles are spread with some uniformity through the rhyolite (Fenner, 1938, Pl. 7, fig. 2).

Many plagioclase phenocrysts from the basalt are isolated in the rhyolite at this stage, some with a thin fringe of black basaltic material still clinging to them, others with no remaining vestments of the basaltic groundmass (Pl. 7, fig. 4). It is significant that these plagioclase xenocrysts show the same albite content ($Ab_{26}An_{74}$ to $Ab_{30}An_{70}$) as the plagioclase phenocrysts of the uncontaminated basalt, and that this albite content is maintained out to the edge of each xenocryst. With a few exceptions, these xenocrysts show no external corrosion effects but retain their characteristic original form. No change in albite content of the phenocryst is apparent along the walls of fractures (Pl. 7, figs. 5, 6).

Some pyroxene phenocrysts at this stage are unaffected, but most of them show marginal alteration to a fine-grained, highly birefringent mineral. Olivine xenocrysts are further altered at their rims to a fine-grained aggregate of a mineral of high refractive index and high birefringence, and in some a little fine-grained hematite has developed.

The abundant quartz and orthoclase phenocrysts at this stage are abnormal in that they are rounded and that the orthoclase phenocrysts show thin rims of turbid material. Augite coronas on the quartz phenocrysts are extremely rare, occurring only in rudimentary form where the phenocrysts lie in basic clots.

Interesting contamination effects are seen in some of the glassy phases of the complex, where numerous tiny needles of a mineral of high refractive index are frequently scattered in a fluxion arrangement through the brown glass. Extinction angles of these needles range up to 45°, suggesting that they are pyroxene. In specimens where the needles are not parallel to each other, a needle or group of needles may occupy a zone in which the glass groundmass is entirely free of the dark particles. It is evident that here the needles have grown at the expense of the dark particles, and, since pyroxenes are not normally present in the groundmass of the rhyolite, it is concluded that they indicate a basification of the rhyolite lava by the basalt.

The description this far has been in the order of increasing contamination of basalt by rhyolitic material and has shown that, as the rhyolitic end member is approached, the large quartz and orthoclase grains increase in abundance and decrease in degree of alteration. To emphasize the correlation between the amount of basaltic material and the frequency of occurrence as well as the degree of alteration of these quartz and orthoclase phenocrysts, these factors will next be discussed in the order of their occurrence in rocks of progressively greater basalt content.

Quartz phenocrysts in the slightly contaminated rhyolite show only rounding and embayment (Pl. 6, fig. 1). With greater contamination by the basalt many quartz grains carry coronas of augite crystals and glass (Pl. 6, figs. 3, 4). In rocks where basaltic material predominates, the coronas of augite and glass around quartz grains are wider, having, in some cases, a greater width than the diameter of the quartz grain itself (Pl. 6, fig. 6). The end stages of this progressive destruction of quartz xenocrysts are represented by the patches of small augite crystals in a matrix of glass and poorly crystallized feldspar (Pl. 6, fig. 5).

Orthoclase phenocrysts in the slightly contaminated rhyolite are rounded, and the outer margins of many of the grains have a minutely wrinkled appearance. With greater contamination by the basalt this wrinkled margin widens, and it appears to be orthoclase of somewhat smaller optical angle than that of the central clear portion (Pl. 5, fig. 3). Favorably oriented, this wrinkled or "fingerprint" texture is seen to be caused by thin stringers of glass lying along the cleavage planes of the orthoclase (Pl. 5, fig. 4). The outline of the phenocryst viewed in some orientations is serrated, while in other orientations the glass formed along the two cleavages separates the orthoclase crystal into tiny rectangular blocks (Pl. 5, fig. 5). The alteration rims of some grains are partly broken away from the clear central portions. In dominantly basaltic rock, grains composed entirely of the wrinkled material are more common than those having cores of clear orthoclase (Pl. 5, fig. 6). The orthoclase xenocrysts

in this relic or "ghost" state are found in specimens in which most of the quartz xenocrysts still retain cores of unaffected quartz within the coronas of augite and glass.

The effect of rhyolitic contamination on the basalt is thus immediately discernible in the change in composition of groundmass feldspar and the decrease in grain size. The phenocrysts of the basalt are affected but slightly until the contamination is far advanced. The effect of basaltic contamination on the groundmass of the rhyolite can only be followed to an intermediate stage, for the rhyolitic material as such then loses its identity. The effect of progressive basaltic contamination on the quartz and orthoclase phenocrysts of the rhyolite depends roughly on the relative amounts of basaltic material with which they are associated. The final stage, where the rock is predominantly basalt, is their complete reconstitution as "ghost" remnants.

The progressive variations outlined above are observable in almost any series of specimens collected along lines across gradational contacts between the rhyolitic and basaltic portions of the complex and, likewise, across apparently sharp contacts. Effects of rhyolitic contamination on the basalt can be observed several yards from any visible or probable contact with rhyolitic rock. The presence of quartz and orthoclase xenocrysts and "ghosts" in the basalt at points far from the rhyolite is considered highly significant. The fact that their representatives can be traced back step by step through the xenocrysts of the less contaminated basalts to the phenocrysts of the normal rhyolite is regarded by the writer as particularly damaging to the hypothesis that they were produced by pneumatolytic action (Fenner, 1938, p. 1449, 1471).

FORMATIONS ASSOCIATED WITH THE COMPLEX

Meadow rhyolitic tuff.—The Meadow rhyolitic tuff is made up of small fragments and shards of glass (in part devitrified), crystals of quartz and orthoclase, a few crystals of sodic plagioclase, and irregularly distributed fragments of foreign derivation (Pl. 8, fig. 6). The accidental material includes fragments of arkose, sandstone, andesite, and various types of basalt, some of which have the characteristics of the early andesites and basalts of Yellowstone Park.

Castle rhyolite.—The groundmass of the Castle rhyolite ranges from a glassy to a micrographic or diffuse spherulitic texture. A typical, though not invariable, feature of this rock is the tiny imbricated shreds of glass or finely crystalline material in the groundmass (Pl. 8, fig. 5). The shreds are parallel and bend as a group around phenocrysts and foreign masses, giving an appearance of the texture produced by viscous flow in a nonhomogeneous material. The ends of most of the shreds are tapered or frayed, and in some cases the shredded portion covers only a narrow band in the rock. The irregularities in distribution of the shredded portions and in the degree of their welding or cementation may account in part for the difference in resistance to erosion of adjacent layers in a cliff face (Pl. 4, fig. 1).

Phenocrysts of the Castle rhyolite are chiefly quartz and orthoclase, most of them idiomorphic, ranging up to 0.3 mm. in diameter.

Sheepeaters basalt.—The even-grained Sheepeaters basalt already has been contrasted with the porphyritic Cataract basalt of the rhyolite-basalt complex. The

texture of the Sheepeaters basalt is intergranular with plagioclase, olivine, and pyroxene making up its constituents (Pl. 8, fig. 4). The plagioclase ranges in length up to 0.3 mm. and in composition from $Ab_{35}An_{65}$ to $Ab_{31}An_{69}$, although this range may be greater in a single crystal. The composition of one olivine grain, determined by measurement of its optic angle, was $For_{76}Fay_{24}$.

Elkhorn basalt.—The groundmass of the Elkhorn basalt is high in granular pyroxenes lying in a network of lath-shaped calcic plagioclase (Pl. 8, fig. 3). Phenocrysts are chiefly plagioclase and olivine. The plagioclase phenocrysts are elongate with ragged terminations, and their composition is about $Ab_{30}An_{70}$, determined by extinction angles. The olivine phenocrysts commonly show hematite as one alteration product.

ORIGIN OF THE COMPLEX

GENERAL CONSIDERATIONS

In the Gardiner River mass two features have been taken by earlier investigators (Iddings, 1899, p. 430–432; Fenner, 1938) as *a priori* evidences that the basalt had been erupted, solidified, and cooled before the eruption of the rhyolite. These features are (1) the readily apparent invasion of the basalt by rhyolite dikes, and (2) the isolation of basalt xenoliths in the rhyolite.

The writer has endeavored to approach this problem from as fundamental a viewpoint as possible and, in doing so, has concluded that: (1) these intrusive features do not necessarily point to an origin by assimilation of already cold basalt, and (2) some features throw doubt on the possibility of such an origin here, while (3) still other features point directly to a mode of origin by mingling of contemporaneous rhyolitic and basaltic lavas.

INTRUSION OF BASALT BY RHYOLITE

The reliability of intrusive and xenolithic structures in determination of relative ages of rocks in the normal case cannot be doubted. A rhyolite-basalt contact, for instance, along which are found rhyolite dikes cutting the basalt and basalt xenoliths in the rhyolite, ordinarily would call for but one interpretation—that the basalt is older than the rhyolite. Such structures in the Gardiner River complex were so interpreted by Iddings (1899, p. 432).

The possibility cannot be overlooked, however, that essentially the same structures would result if the solidification of the basalt had not taken place until the actual moment of contact with the rhyolite. In such a case the geologic ages of the two rocks would be identical, and the ordinary interpretation of intrusive structures would be incorrect. Since there is no indication in these structures of the length of time the basalt might have been solid previous to its invasion by the rhyolite, we can draw only one absolutely safe conclusion: When the two materials met, the rhyolite behaved as a liquid and the basalt as a solid.

A basis for the belief that molten basaltic lava would solidify upon contact with molten rhyolitic lava is furnished by differences in the normal crystallizing temperatures of the two types of lavas. Larsen (1929), in evaluating accumulated data on

the temperatures of magmas, has concluded that the temperatures of most crystallizing rhyolitic magmas are probably not far from 600° to 700°C., while those of most crystallizing basaltic magmas are probably close to 800° to 900°C. It is worth while, therefore, to examine the possibility that the cooling effect of the rhyolitic lava was sufficient to cause solidification of the basaltic lava.

The abundant phenocrysts in each member of the Gardiner River complex indicate that both these lavas had started to crystallize before extrusion and consequently were at their respective normal crystallizing temperatures; the phenocrysts of uncontaminated rhyolite and those of the uncontaminated basalt are appropriate to their respective rock types and show no abnormal corrosion features. On the basis of Larsen's conclusions, then, the temperature of the rhyolitic lava was probably 100°–300° lower than that of the basalt.

The consistent production of basaltic xenoliths rather than rhyolitic xenoliths would be a natural result of the mutual effects of the two intermingling lavas. Since the lavas would be in motion, the contact between the two moving liquids could not be stable, and small masses of one liquid would be dragged into the main masses of the other liquid by the turbulent movement. A small mass of basaltic liquid, isolated in rhyolitic liquid, would immediately give off some of its heat to the cooler surrounding material. As the basaltic liquid was already crystallizing, any sudden lowering of temperature should accelerate the speed of crystallization so that it would almost immediately become a fine-grained mush, or even a complete solid. In either case its behavior would be more like a solid than a liquid while involved in movement with the surrounding viscous rhyolitic liquid.

In the initial stages of mushiness, the basaltic mass would have been drawn out into elongated and rounded xenoliths. As the coherency of the basaltic material became greater and fragmentation due to flowage of the enclosing viscous rhyolitic lava continued, angular forms also would begin to appear.

Considering now the small blebs of rhyolitic liquid dragged into the main mass of basaltic lava, no parallel tendency toward the formation of rhyolitic xenoliths is discovered. Rather, a rhyolitic bleb, quickly heated to the temperature of the surrounding basaltic material, would cease crystallizing and would dissolve in the basaltic liquid. The resorption of the already formed quartz and orthoclase phenocrysts would proceed more slowly, and these crystals might persist as xenocrysts for some time after complete assimilation of the rhyolitic liquid.

The formation of rhyolite dikes, as well as the prevention of the formation of basaltic dikes, would be controlled by this same factor of difference in temperature between the two lavas. Some thin stringers of rhyolitic material would be drawn out and diffuse partially or completely into the basaltic lava. At later stages the absorption of the rhyolitic material would be handicapped, and a plexus of rhyolitic veinlets would result. If stringers of basaltic material were drawn into the moving rhyolitic liquid, however, they would quickly solidify and break up to form xenoliths.

With more rhyolitic liquid dragged into the basaltic lava, incomplete absorption of the rhyolite before final consolidation would lead to the production of a visible streaky effect. Structures similar to that of "marble cake" would be expected where the proportions of rhyolitic liquid and basaltic liquid in the mixture were correct (Fenner,

1938, Pl. 8, figs. 4, 6). In all cases, the tendency of diffusion to equalize composition would be handicapped by the high viscosities of the mixtures and by the relatively short time available before final consolidation of the composite flow.

Following the process still further, wider and more persistent rhyolitic dikes would be formed at later stages in the cooling history of the composite rhyolite-basalt mass. The basaltic lava would first be mushy or solid where it was in contact with the cooler rhyolitic lava. As the whole mass cooled, the mushy or solid shell of the basaltic portions in contact with the rhyolitic lava would thicken, but the rhyolitic lava would remain essentially liquid. Continued movement of the composite lava mass would result in the fracturing of the basaltic shell, permitting the molten rhyolite to enter into it and to form dikes of significant width and length. Angular xenoliths, derived previously or torn from the not yet completely rigid basaltic material on the walls of the fissure, would be carried along by the rhyolitic liquid.

This process, as outlined, would point to a consistent tendency toward the production of basaltic xenoliths and rhyolitic dikes during the mingling of molten basaltic lava and molten rhyolitic lava. It is a conception much at variance with the commonly expressed opinion that the mixing of two lavas or magmas should give rise throughout to a texture similar to that of "marble cake," with neither material showing a consistent tendency to form islands of xenoliths. The validity of the process suggested here rests on the ability of the rhyolitic liquid to exert enough cooling effect to induce partial solidification of the basalt and upon the ability of the rhyolitic lava to remain liquid at an intermediate stage in the cooling of the composite mass after large portions of the basalt have already solidified. If this process is accepted as feasible, the presence of basaltic xenoliths in the rhyolite and of rhyolitic dikes in the basalt of an igneous complex of this type cannot be regarded as conclusive proof that the basalt is older.

FORM OF CONTACT BETWEEN RHYOLITE AND BASALT

Fenner (1938, p. 1445, 1446) has pointed to the apparent situation of the rhyolite in the bottom of a V-shaped trough in the basalt as a further indication of the greater age of the basalt. This "valley" in basalt, approximately in the position of the present valley of Gardiner River, he considers to have guided the flow of the rhyolite as it actively attacked the basaltic walls and bottom. However, the crude trough shape of the contact could be produced if the rhyolite and basalt met as two moving flows or if the rhyolite were erupted through a basaltic flow that was moving across the First Cataract area. If a lobe of a rhyolitic flow encountered a basaltic flow, the rhyolitic flow would tend to ride over the basaltic flow because of the rhyolite's lesser density, but it would be prevented from spreading thinly over the basalt because of its high viscosity. It is also possible that the eruption of the rhyolite took place at First Cataract, since the base of the rhyolite is not exposed over much of the area and there is nothing in the flow structures of the inner portion of the complex to refute such a suggestion. In either case, the resulting form of the contact, while irregular in detail, would be roughly trough-shaped.

The relative positions of the rhyolitic and basaltic masses as shown by the maps (Pls. 1, 2) imply that the rhyolitic material lies against slopes of basalt, with the

original rims of the basalt valley exposed at higher elevations than the rhyolitic material. In several instances, however, rhyolite apparently lies underneath basalt rather than above it. At the eastern end of the complex (Fenner, 1938, Pl. 6, fig. 1) rhyolite underlies a mass of greatly altered basalt 200 feet long. The latter case Fenner attributed to undercutting of the basalt cliff by a stream of molten rhyolite. However, the flow planes in the rhyolite plunge directly into the cliff below the basalt. If the basalt had been undercut by the rhyolite, or by normal subaerial erosion, it is hardly conceivable that the undercutting could have proceeded into the basalt for more than a few feet. In that event, some reflection of the embayed basalt surface should be seen in the configuration of the flow planes through that 10 feet of rhyolitic material exposed below the basalt mass. This effect, however, does not seem to be revealed, for the flow planes in the rhyolitic material generally parallel the lower contact of the basalt mass throughout the exposure.

Other departures from normal erosional forms are represented by the wider dikes of rhyolite cutting massive basalt (the "pseudo dikes" of Fenner, 1938, Pl. 4, fig. 3; Pl. 5, fig. 2), which, according to Fenner, might be rhyolitic fillings of clefts in the basaltic wall of the erosional valley. Strong development of flow planes in the rhyolite parallel to the walls of the dike suggests, however, that the movement of the rhyolitic material was much stronger than the type of movement involved in the simple filling of a vertical cleft in a valley wall. In another case a horizontal mass of basalt is underlain by 4 feet of rhyolite (Fenner, 1938, p. 1459, Pl. 4, fig. 3). The configuration of the basalt surface here is very difficult to explain as a previous overhanging ledge of basalt or as due to undercutting of a basalt cliff by corrosive rhyolitic lava.

Therefore, if it be assumed that a stream valley had been eroded into the basalt previous to the advent of the rhyolitic lava flow, it seems necessary to draw upon abnormal circumstances and processes to explain the irregularities cited above. On the other hand, if it be assumed that both materials came together as liquid lavas, one of which would become essentially solid before the other, the shape of the rhyolite-basalt contact is not extraordinary. Under those circumstances the contact is expected to be irregular, involving many examples of rhyolitic dikes in various orientations, horizontal as well as vertical.

RELATION OF THE COMPLEX TO OTHER FORMATIONS

The discussion so far has attempted to show that neither the intrusional nature of the rhyolite nor the form of the rhyolite-basalt contact demonstrates conclusively that the rhyolite is younger or the same age as the basalt. If the arguments put forth are acceptable, the problem of the origin of the complex must then require further evidence before a final decision can be made. If the rhyolitic lava flow had a profound effect on a previously existing basalt, it is of interest to examine the effect of the rhyolitic lava upon other formations with which it came in contact.

One older formation, the Meadow rhyolitic tuff, outcrops on the banks of the meadow immediately east of the complex; it is considered pre-complex in age on the basis of its field relations. The glassy chilled selvages of the rhyolitic portions of the complex apparently lie against this pre-existing surface on tuff.

No glassy selvages of the rhyolitic portions of the complex are found, however, against the basaltic masses of the complex. Instead, the rhyolitic and granophyric textures persist right up to the actual contact of the rhyolitic material with the basalt. If the basalt, as well as the tuff, had been present as a cold rock mass before the arrival of the rhyolite, the question remains: Why did the rhyolite not chill to a glass against the basalt as well as against the tuff?

Three alternatives that might account for the divergent behavior of the rhyolitic lava are: (1) The basalt, though cool, might have been unable to extract sufficient heat from the rhyolitic lava to cause chilling, (2) exothermic reactions at the rhyolite-basalt contact might have produced sufficient heat to prevent chilling, or (3) the basalt may itself have already been hot at the time of contact.

(1) The thermal conductivity of basalt according to the International Critical Tables is 20 joules per cubic centimeter at 20°C. Although no data are given for the conductivity of tuff, the conductivities of a substance such as pumice, which is reasonably similar in composition and texture, should be acceptable for comparison. The thermal conductivities of two samples of pumice are given as 0.92 and 1.86 joules per cubic centimeter at 20°C. It is thus apparent that the ability of basalt to conduct heat from the rhyolitic lava (and thereby to chill it) would be much greater than that of pumice and doubtless also of tuff. As for relative heat capacities, the difference between the heat capacity of basalt and of the average heat capacity of the constituents of tuff, including that of air, is apparently too small to have a significant effect (Daly, 1933, p. 63; Hodgeman and Holmes, 1940, p. 1732). Therefore, the amount of heat absorbed from a molten rhyolite by a cold basalt, per degree rise in temperature of the basalt, should not be less than that extracted from a molten rhyolite by a tuff. Taking into account the greater thermal conductivity of the basalt, the rate of extraction of heat from the molten rhyolite—*i.e.*, the rate of cooling of the molten rhyolite—should be greater at the basalt contact than at the tuff contact. If, then, this factor had been operative, it would have tended to produce a wider glassy margin at the basalt contact than at the tuff contact. This possibility is therefore regarded as inadequate to account for the existing reverse relation.

(2) Examination of the microscopic characters of the contact zone furnishes pertinent information about heat-producing reactions at the contact. The lack of any significant amount of reaction products in the contaminated basalt has been pointed out by Fenner, who says (1938, p. 1472):

> . . . many instances of reaction products should be visible. The equilibrium among the minerals of the basalt should have been altered, and reactions consequent upon solution of olivine and of pyroxene (and perhaps of magnetite) in the siliceous liquid should have given pyroxene of changed composition or some new mineral, such as biotite or hornblende. On the other hand, if a separate vapor phase was the medium, it might well have escaped entirely, and the residual liquid need not contain reaction minerals.
>
> "The results actually found are as follows. In the transformed rocks the olivines are plainly the corroded remnants of original olivines, and the pyroxenes are the corroded remnants of original pyroxenes. No new pyroxene (or only a very small quantity) is recognizable, and biotite and hornblende are present in insignificant amount."

To explain the undoubted contamination of the apparently solid basalt at a distance from the rhyolite, Fenner proposed the agency of tenuous rhyolitic vapors. Because no significant amount of reaction products with the ferromagnesian minerals are seen

in the resulting contaminated rock, he concludes that these rhyolitic vapors, driven onward by pressure originating within the rhyolitic flow, carried the reaction products on through the solid basalt to some now unseen destination outside.

At this point, however, a critical inconsistency becomes apparent. If the vapors escaped with their load to an outside destination through the solid basalt wall of the postulated valley, we must also consider the progress of the vapors that were penetrating the larger basalt xenoliths from all sides. It is only reasonable in this circumstance that the same process that produced the contamination of the basalt wall rock must have acted to produce the contamination of the basalt xenoliths, since the progressive changes seen are identical. The reaction products would then have been carried to the center of the xenoliths from all sides and, having no further escape, would have been deposited there. When we look for these reaction products in the central portions of the xenoliths, however, we find no trace of them but only a gradation into unaltered basalt.

The lack of reaction products in this one critical situation must make it doubtful that there ever existed any significant amount of reaction products in the sense of those formed between rhyolitic liquid or vapor and solid basalt. The possibility of there having been any significant amount of reaction, exothermic or otherwise, between the two materials therefore must be equally doubtful.

(3) There remains the third alternative—that the basalt was already hot when the rhyolite came in contact with it, and its merits will receive detailed consideration later.

COMPOSITIONAL RELATIONS BETWEEN PHENOCRYSTS AND GROUNDMASS OF CONTAMINATED BASALT

While the nature of the groundmass ferromagnesian minerals of the contaminated basalts could not readily be determined optically because of their small size and imperfect crystallization, the groundmass plagioclases could be determined, and they revealed an albite content that is roughly a function of the degree of contamination. That these groundmass feldspars are no ordinary reaction products formed in solid rock is shown, however, by the lack of a parallel increase in the albite content of the plagioclase phenocrysts of the contaminated basalts. Much significance is attached to this feature, for it indicates that the vulnerability of the groundmass at the time of contact must have been much greater than that of the phenocrysts.

If the basalt were a previously consolidated rock, the rhyolitic vapors, in order to contaminate basalt several yards from the rhyolite, would have had to penetrate and soak through the solid basalt over that distance. It seems inconceivable that intimate penetration of the rock mass by vapors that had such a marked effect on the composition of the groundmass feldspars could, at the same time, have had no effect on the phenocrysts.

The wide difference in vulnerability between the groundmass and the phenocrysts must have some cause other than that of original differences in composition, for in the normal Cataract basalt the difference in albite content between groundmass plagioclase and phenocryst plagioclase is no more than 5 per cent. Neither does it seem possible that differences in relative amounts of exposed surface could be a cause, for

with complete permeation of the solid rock by these vapors there is no reason to expect that the vapors would avoid the incipient cleavages and fractures in the phenocrysts.

Contamination of the basalt by pneumatolytic vapors would therefore seem doubtful since no trace of their effect is seen on the plagioclase phenocrysts. The remarkable lack of chemical or physical alteration of these phenocrysts, even where isolated as xenocrysts in the rhyolite, is unexplainable under the assumption of a previously solid porphyritic basalt. It is just this feature that would be expected if the basalt had been a hot "porphyritic" liquid or mush when invaded by the rhyolite.

XENOCRYSTS OF QUARTZ AND ORTHOCLASE IN CONTAMINATED BASALT

The presence of large grains of quartz and orthoclase in the contaminated basalt many feet from the rhyolite is most difficult to explain on the assumption of an initially cool and solid basalt.

There can be little doubt that quartz and orthoclase grains were derived from the rhyolite. Though visibly affected in their outer portions, they are otherwise identical with the quartz and orthoclase phenocrysts of the rhyolite, and their numbers increase as the amount of rhyolitic material in the contaminated basalt increases. They show the same specific variations in habits found in the rhyolite phenocrysts—namely, occasional grains with nuclei of clear sodic plagioclase inside rims of clear orthoclase, and grains with nuclei of patchy and wavy extinction inside rims of clear orthoclase. This, together with the fact that these greatly reconstituted grains fall in a definite alteration series, can be regarded as nearly conclusive evidence that they were originally phenocrysts in the molten rhyolite and did not grow in place, as has been proposed for some of them by Fenner (1938, p. 1449, 1471). It should not be expected that all the quartz and orthoclase xenocrysts in the contaminated basalt would carry such reaction rims for in many cases the movement of the surrounding liquid would remove the alteration material as fast as it was formed, leaving smaller grains.

The alteration series described above is almost identical with similar progressive effects described by Lacroix (1893, p. 18, figs. 1, 2; Pl. I, fig. 1) in quartz and orthoclase xenocrysts and xenoliths isolated in basaltic flows. According to Lacroix, fusion of the peripheries of quartz crystals is followed by growth of augite projecting into this vitreous zone from the basalt, or distributed at random in the vitreous zone. Alkali feldspar xenocrysts isolated in the basalts melt initially along the cleavages, leading to serrate edges or to a rectangular mosaic, depending upon the section viewed (Lacroix, 1893, p. 56; Pl. II, figs. 1, 2; Pl. III, fig. 12. *Compare* this paper, Pls. 5, 6).

If we assume that the basalt of the Gardiner River complex was solid at the time of attack by the rhyolite, some method must be postulated for the entrance of these xenocrysts of quartz and orthoclase into the basalt. It might be suggested that they gained access to basaltic xenoliths after a preliminary melting and softening of the xenoliths which allowed them to be pressed into the resulting mush. Such a melting by the rhyolite could not have been effective, however, over the distance of several yards into the main mass of basalt, where these xenocrysts are also present. If, however, we assume the basalt to have been molten or mushy when the rhyolite en-

countered it, the bodily introduction of the quartz and orthoclase phenocrysts into the basalt would come about quite naturally as a result of the movements of the two materials.

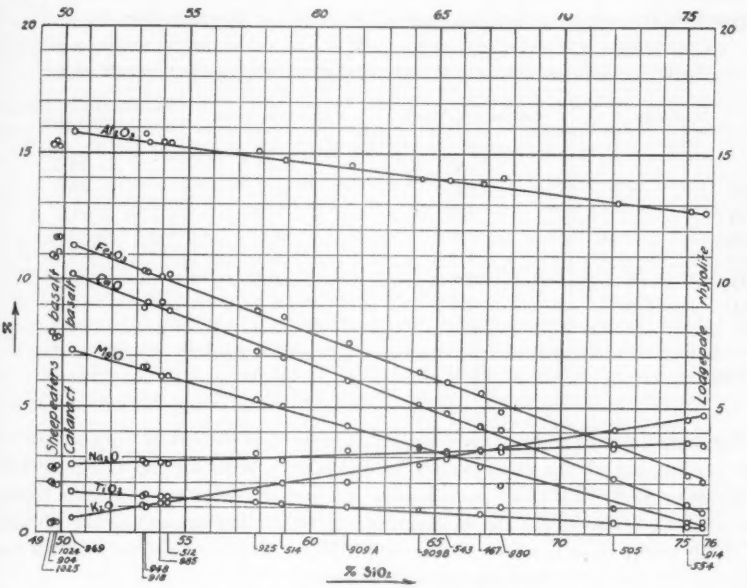


FIGURE 2.—*Chemical variation diagram*

Gardiner River rocks (Modified from Fenner, 1938, p. 1476)

CHEMICAL RELATIONSHIPS OF THE COMPLEX

On the assumption of a solid and cold basalt, the strictly rectilinear variation in chemical composition of the rock series of the complex would require the assignment of extraordinary powers to rhyolitic vapors proposed as the agents in the modification of the solid basalt. These powers would include the ability of the vapors (1) to convey rhyolitic constituents into the basalt in the exact proportions in which they had occurred in the rhyolite, (2) to remove, at the same time, basaltic constituents in the exact proportions in which they occur in the basalt, and (3) to accomplish these things without significant increase in the volume of the basaltic mass (Fenner, 1938, p. 1471, 1472). A critical examination should be made here of the details of the possible mechanism by which this process might have proceeded.

The variation diagram published by Fenner (1938, p. 1476) illustrating the chemical relations is reproduced here with added notations (Fig. 2). The three analyses of specimens lowest in silica (Nos. 904, 1024, 1025) taken by Fenner (1938, p. 1450) from the "rim basalt" already have been assigned in this paper to Sheepeaters basalt, a younger flow, which did not participate in the formation of the complex. Fenner's specimen No. 969, however, was taken from the Cataract basalt at Second Cataract

and represents the uncontaminated basalt of the complex. Specimen No. 914 shows the highest silica content and probably represents the essentially uncontaminated Lodgepole rhyolite of the complex. The names of these formations have been superimposed on the corresponding analyses of the variation diagram (Fig. 2).

TABLE 1.—*Gains and losses during supposed modification of basalt at constant volume*

Constituent	Gms. entering basalt	Gms. leaving basalt	Gms. net change of solubility in fluid agent
SiO_2	75.6	56.3	−19.3
Al_2O_3	12.6	17.7	+5.1
TiO_2	0.1	1.9	+1.8
Fe_2O_3	2.2	12.8	+10.6
MgO	0.3	8.1	+7.8
CaO	0.8	11.5	+10.7
Na_2O	3.6	3.1	−0.5
K_2O	4.8	0.6	−4.2
Total	100.0	112.0	+12.0

The percentage of silica in each analysed specimen is plotted as the abscissa with the percentages of the other oxides as ordinates. With all the Gardiner River analyses so plotted the points representing the percentages of any one oxide through the series of specimens fall almost exactly on a straight line connecting the extremes. With the exception of the plots of Nos. 904, 1024, and 1025, none of the points is more than 0.5 per cent away from its appropriate line.

If it is assumed that constant volume was maintained in the basalt undergoing modification, as proposed by Fenner, then the addition of a given amount of rhyolitic constituents to the basalt must have been accompanied by the removal of an equal volume of basaltic constituents in the exact proportion in which they had occurred in the basalt.

To illustrate the steps in this supposed process, let us take, as an example, a given mass of basalt of specific gravity 2.9 that is to receive 100 grams of rhyolitic material of specific gravity 2.6. The weight of the equal volume of basaltic material that must be removed to make room for the rhyolitic material would be proportional to the relative specific gravities and should be equal, therefore, to 112 grams of basaltic material. Using the plots of the analyses of No. 914 (uncontaminated rhyolite) and No. 969 (uncontaminated basalt) as a basis, the gains and losses in each constituent of the basalt have been calculated (Table 1).

In the first column the 100 grams of rhyolitic material entering the basalt are broken down into specific constituents, while in the second column the same is done for the 112 grams of basaltic material being removed. The last column gives the net gains or losses of each constituent in the modified basalt, and this could be regarded as a measure of the changes in solubility of the constituents being carried by the vapor agents.

Comparing the figures in adjacent columns, it is seen that as these vapors worked

their way through the rock their ability to carry the constituents must have changed radically, for upon finally leaving the rock they were carrying the constituents in entirely different and special proportions. The vapors brought in 75.6 grams of silica, but this same amount of vapors took away only 56.3 grams of silica; that is, 19.3 grams of silica were left in the basalt. Likewise, while the vapors brought in 12.6 grams of alumina, they removed 17.7 grams, a net removal of 5.1 grams of alumina from the basalt. Titania, iron oxide, and the rest could be described in a similar manner. The average effect was that 12 more grams of material were removed than were brought in.

How would the solubilities of these constituents in the vapors change so markedly during transit through the solid basalt? Would they have changed gradually or at some particular instant? It does not seem possible that at one instant the vapors could have been carrying these constituents in one proportion and at the next instant have been carrying them in an entirely different proportion. However, if the solubilities changed gradually they must all have changed proportionately and at the same rate. Otherwise, the rectilinear proportions of the constituents could not be maintained in chemical analyses of specimens of contaminated rocks picked at random. If the changes in solubility did not proceed in this very special manner, the plots of the analyses would not fall on straight lines but would be scattered at random or distributed along a curve.

These delicately conditioned and possibly mutually independent effects, moreover, would have been taking place during a general cooling of the vapors as they made their way farther and farther into the cold basalt. Thus, with the gradual drop in temperature, the magnesia, lime, iron oxide, and alumina would become relatively more soluble than the silica, potash, and soda, while at the same time the total weight of material in solution would have been increasing. If such a relative increase in solubility of basic constituents really existed, it would be a reversal of the trend well displayed in earlier magmatic stages during the crystallization of gabbroic liquid, where these constituents become relatively less soluble with decreasing temperatures.

The postulate that vapors were driven into the solid basalt from the rhyolite lava requires that a pressure gradient must have been present through the basalt. The process of modification must have ceased at some time, yet where are the basaltic constituents (the reaction products) that were being removed when the pressure and temperature gradients degenerated? Admittedly some should have been trapped during the dying stages of the activity of these vapors, but none are present, as has already been noted.

Therefore, the chemical relationships of the rocks of the complex would require radical readjustments in the present concepts of the mechanism of physico-chemical processes if these relationships were established in solid rock by pneumatolytic vapors. These readjustments in concept should certainly be given most serious consideration if all other observed features in the complex supported the proposal that pneumatolytic vapors contaminated the solid basalt. Since other characteristics of the complex clash sharply with this proposal, however, we are led to doubt that the pneumatolytic vapors were effective.

HYPOTHESIS OF SIMULTANEOUS ERUPTION OF DIVERSE MAGMAS

A critical re-examination of the rhyolite-basalt complex on Gardiner River shows that each feature observed fits easily and logically into a hypothesis based on the nearly simultaneous extrusion of the rhyolite and basalt as liquid lavas. This hypothesis provides a basalt that was neither rigidly solid nor cold but was incoherent to varying degrees and possessed a high temperature.

Evidence has already been presented for the belief that an encounter between molten rhyolite lava and molten basaltic lava would produce basaltic xenoliths in rhyolite and rhyolitic dikes in basalt, as well as the intimate "marble-cake" type of mixtures.

The visible parts of the rhyolite-basalt contact suggest that the rhyolite lies in a rough V-shaped trough in the basalt. The irregularities of that trough, as has been already pointed out, are such that they could hardly have been developed by subaerial erosion. In some cases, such as that of the horizontal basalt layer both overlain and underlain by rhyolite, no reasonable explanation based on an erosion valley in previously existing basalt or on corrosion by molten rhyolitic lava appears to be available. All these irregularities in the shape of the contact are to be expected, however, if the complex had originated upon the meeting of two simultaneously erupted lava flows. This includes the major irregularity of the apparent V shape of the rhyolite-basalt contact, for the rhyolite lava, being less dense, would tend to override the moving basalt lava.

A chill phase of the rhyolite lava against the basalt would not form if the basalt had not yet cooled sufficiently following its extrusion, but the chill phase should form against older formations, such as the Meadow tuff.

Contamination of the basalt by the rhyolite would have been immediately effective in the still liquid or mushy portions of the basaltic lava, and groundmass minerals subsequently crystallizing from the contaminated liquid would have intermediate compositions. These minerals of intermediate composition would not be "reaction products" in the sense of having been formed by reaction between rhyolitic lava or vapors and solid basalt. Phenocrysts of bytownite, pyroxene, and olivine, already present as crystalline phases in the basaltic liquid, would have been relatively less vulnerable to attack and alteration by the contaminated basaltic liquid. This is especially indicated in view of the relatively short time available before final consolidation set in. The lowering of the temperature of the liquid due to incorporation of the cooler rhyolitic liquid and the possible endothermic effect of the actual mixing of the diverse liquids (Bowen, 1928, p. 180) may have been an added handicap to the ready assimilation of the already formed mafic phenocrysts.

Xenocrysts of quartz and orthoclase from the rhyolitic lava could quite easily have entered the basalt liquid or mush during mutual movement of the two lavas, and, with continued movement, many would be removed far from the major contact with the rhyolite proper. The small portions of rhyolitic liquid that accompanied them into the basalt lava would be relatively quickly absorbed into the hot basaltic liquid. This would leave the quartz and orthoclase grains apparently isolated in the basalt with no other indication of contamination than the slightly less mafic composition of the groundmass minerals.

Alteration of the xenocrysts of quartz and orthoclase would be expected to proceed

more rapidly than the alteration of the xenocrysts of bytownite, pyroxene, and olivine in analogous situations in the rhyolite. The quartz and orthoclase grains would be in a liquid much hotter than the rhyolitic liquid from which they were derived and with which they had been essentially in equilibrium. The slightly greater rate of reduction of orthoclase xenocrysts to "ghost" remnants compared to that of the quartz xenocrysts would be caused, in part at least, by the abundant planes of cleavage in the orthoclase along which melting could proceed.

Since rectilinear variation in chemical composition is characteristic of mechanical mixtures, this feature would be expected in the mingling of diverse lavas. Here the mixing is between rhyolitic liquid and basaltic liquid and also between rhyolitic liquid and mushy basaltic material. The rigid adherence of the analyses to a straight-line relationship in a variation diagram would be a natural result of such mechanical mixing, whether the materials were liquid or solid. The very simplicity of this hypothesis, compared with the complex mechanisms required by the hypothesis of a previously solid basalt, would be an argument in its favor in ordinary circumstances.

It should be necessary, in connection with any proposal of the mixing of two molten lavas of divergent composition, to account for the circumstances of their close association in time and place. The general requirements for such a relationship would seem to include: (1) a pair of subterranean magma bodies, one rhyolitic and the other basaltic in composition, in relative proximity to each other, and (2) simultaneous tapping of the two magmas with resulting eruptions at orifices not far apart. Because the possibility of such relationships has not usually been drawn on in the literature of volcanic problems, it will be discussed in the remainder of this paper and applied to the Gardiner River occurrence.

Examples of extreme proximity of volcanic orifices is furnished by the "composite intrusions" in Scotland and other regions (Richey and Thomas, 1930, p. 62). Magmas of widely differing composition have apparently occupied the same fissure at nearly the same time to form dikes, sills, or sheets, some of which were undoubtedly feeders for surface eruptions. These intrusions possess many characteristics in common with the rhyolite-basalt complex at Gardiner River. A partial bibliography of the Scottish and other occurrences, descriptions of which show striking similarities to the characteristics of the Gardiner River complex, is appended to this paper.

There are few modern examples of simultaneous eruption of divergent materials from near-by orifices, but the eruptive centers of Vulcano and Stromboli in the Lipari Islands might be cited as one illustration. Vulcano has erupted rhyolitic material at intervals since 1800, while Stromboli, 30 miles north, has been erupting basaltic material almost continuously over the same period. Flows from vents 30 miles apart might, under favorable conditions of topography and supply of lava, meet and mix with each other. Although obviously in the Lipari group neither topography nor supply has been favorable for this encounter, basalt lava, because of its high fluidity and density, can move great distances on a low gradient. The Veidivatnahraun flow of Iceland is traceable more than 90 miles, the southern 30 miles of which slopes only 0°05′. The Frambuni flow in Iceland is 60 miles in length, sloping 0°41′ to 0°30′ (Thoroddsen, 1906, p. 144). A basalt flow in Williams Canyon, Arizona, is traceable for more than 13 miles from its point of origin (Daly, 1933, p. 139).

Suggestions of the nearly simultaneous availability of rhyolitic and basaltic lavas

are found in the Yellowstone region itself. Iddings (1896, p. 614) points to the close association of rhyolite and basalt in Yellowstone Park and to the lack of evidence of long periods of erosion separating some of the eruptions. "Basalt streams in rare instances lie between rhyolite in such a way as to prove their comparative contemporaneity."

One such case might be at Collonade Falls on Bechler River (Iddings, 1899, p. 376) where a basalt flow is both underlain and overlain by rhyolite with the scoriaceous top of the basalt evidently uneroded, implying that there was not time for erosion to become effective before the outpouring of the upper rhyolite. Another example of interleaving of rhyolite and basalt flows on the east wall of Gardiner River canyon several miles downstream from Sevenmile Bridge is cited by Howard (1937).

It is interesting and perhaps highly significant that these alternating eruptions of rhyolitic and basaltic lavas in Yellowstone Park took place at about the same time as the formation of the rhyolite-basalt complex at First Cataract. This was apparently an interval of transition from mainly rhyolitic eruptions to mainly basaltic eruptions.

These relations suggest that two subcrustal magmas of widely different composition existed at the same time not far from each other in the Yellowstone region and were alternately tapped so that their lavas were successively superimposed.

The possibility of simultaneous tapping of two magma chambers, upon close examination, does not seem remote. There is abundant evidence that many of the basaltic eruptions of the world originate from fissures tapping deep-seated magma chambers, whereas most rhyolitic eruptions originate from chambers relatively shallow in the crust. This would indicate that rhyolitic magma might even lie in a separate chamber more or less directly above chambers of basaltic magma, and that both might be tapped by the same or simultaneously formed fissures with a resultant mixture of the two magmas during eruption. Holmes (1931) offers the additional suggestion that rhyolitic and basaltic magmas could exist in the same chamber, maintaining their identities as a result of gravity stratification. Such a dual magma, he postulates, was the parent of the composite intrusions and hybrids of the Scottish districts.

It can therefore be said that, under special conditions, commingling of lavas can reasonably take place. The Gardiner River area at the time of formation of the complex possessed the required conditions as it was in a transitional period in its volcanic history when both rhyolitic and basaltic magmas were available for simultaneous tapping.

CONCLUSIONS

The views of Iddings and of Fenner that the rhyolite-basalt complex on Gardiner River originated through assimilation of cold basalt by molten rhyolite or its emanations are based mainly on the following points:

(1) The intrusive nature of the rhyolite in contact with the basalt, implying that the basalt is older than the rhyolite.

(2) The position of the rhyolite in a roughly V-shaped trough, implying a previous erosion valley in the basalt.

(3) The gradation in composition of specimens from the basalt to the rhyolite of the complex, implying chemical interaction and assimilation.

These points are not regarded in the present paper as being critical, or as pointing necessarily to an origin by assimilation because:

(1) The intrusive nature of the rhyolite in contact with the basalt would also be expected with the mixing of simultaneously extruded rhyolitic and basaltic lavas. With each at its normal temperature, the rhyolitic lava could exert a cooling and solidifying effect on the basaltic lava, resulting in the production of basaltic xenoliths and, at later stages, in the production of the rhyolitic dikes.

(2) The V-shaped trough in the basalt is only approximate, and the position of the rhyolite in it might have come about through either mode of origin. Some of the irregularities in the shape of the basalt contact show more resemblance to structures developed by interfingering of two divergent fluids than to erosional forms.

(3) The gradation of chemical composition in specimens of the complex is strictly rectilinear and, therefore, is less likely a development of chemical attack and assimilation than of mechanical mixing of the two materials, in liquid or solid form, or both.

(4) Fenner has proposed the agency of vanished rhyolitic emanations for the contamination and assimilation of the supposedly solid basalt and for the removal of the resulting reaction products to points beyond observation in the present exposures. However, large basaltic xenoliths, the central portions of which are not contaminated, contain none of the expected reaction products that would under these conditions have been carried inward by the emanations attacking the xenolith from all sides. Since there seems to have been no other path for their escape, the absence of reaction products there is regarded as throwing doubt on their possible existence at any stage.

Three additional points appear to be unexplainable on the basis of an origin through assimilation and seem to indicate definitely that the basalt was hot and more or less fluid when the rhyolitic lava encountered it:

(5) Examination of the glassy phase of the rhyolite against the country rock has shown it to be a chilled margin. The lack of such a chilled margin against the basaltic portions of the complex can therefore be regarded as evidence that the basaltic portions were hot at the time of contact.

(6) The xenocrysts of quartz and orthoclase in the basalt must originally have been phenocrysts in the rhyolitic lava. Their penetration into the basalt to points several yards away from the rhyolite is regarded as an indication that the basalt was fluid or mushy at the time of contact.

(7) The immediately observable effect of slight contamination on the composition of the groundmass feldspars of the basalt, with no corresponding effect on the phenocrysts of the basalt, even with advanced contamination, is regarded as an indication that the groundmass of the basalt was in an extraordinarily vulnerable chemical condition at the time of contact—*i.e.*, it was liquid.

On the basis of these points, it is concluded that the features of the Gardiner River complex do not apply to the question of the ability of salic vapors to accomplish significant assimilation at ordinary pressures or to a discussion of the relative merits of

the theories of assimilation and gravitational differentiation to account for the production of rock types.

Attention is called to the need for critical examination of the evidences for age relationships in salic-mafic rock associations. The suggestion is made that, where salic igneous rock intimately intrudes mafic igneous rock, the salic rock may be of the same age as the mafic rock, and that supplementary evidence should be found before a final decision concerning the relative ages is made. The suggestion that the two rocks may be of the same age, even though the salic rock intrudes the mafic rock, is based on the normally lower temperature of salic magmas compared to mafic magmas and rests ultimately on the question of the ability of the salic magma to cool the mafic magma sufficiently to induce partial solidification.

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University of Wisconsin, Madison, Wisconsin.

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EXPLANATION OF PLATES

PLATE 5.—PHOTOMICROGRAPHS OF ORTHOCLASE PHENOCRYSTS AND XENOCRYSTS IN CONTAMINATED ROCKS

Figure

1.—Compound phenocryst of orthoclase and plagioclase in rhyolitic obsidian. The plagioclase in the central portion is $Ab_{74}An_{26}$. (Crossed nicols, ×16).

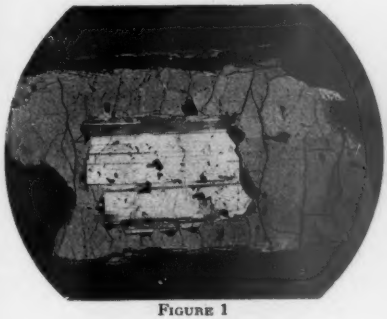
2.—Compound phenocryst of orthoclase and soda orthoclase (?). Note shadowy cross twinning. (Crossed nicols, ×16).

3.—Orthoclase phenocryst in contaminated rhyolite showing initial stage of melting and solution along cleavages. A fragment of a bytownite xenocryst lies to the left of the orthoclase phenocryst. (×16).

4.—Twinned orthoclase in contaminated rhyolite showing attack of crystal along the cleavages and the twinning plane. White V-shaped band is a fracture in the thin section. (×31).

5.—Orthoclase phenocryst in contaminated rhyolite showing rectangular block pattern developed by melting along both sets of cleavages. (×31).

6.—Orthoclase "ghost" xenocryst in contaminated basalt. The "ghost" is composed of innumerable isolated islands of feldspar material, apparently changed in composition from that of the original orthoclase. (×31).



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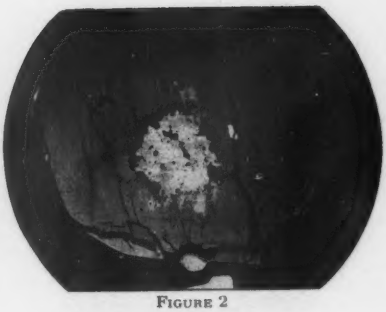

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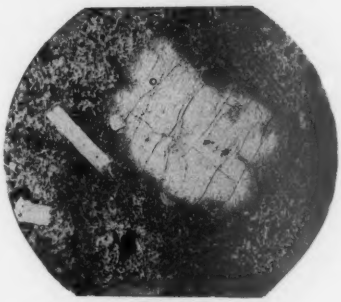

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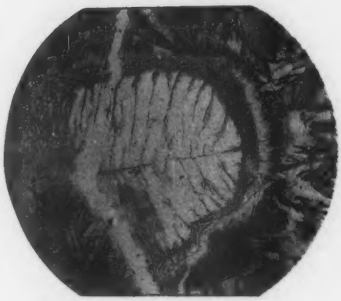

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FIGURE 5

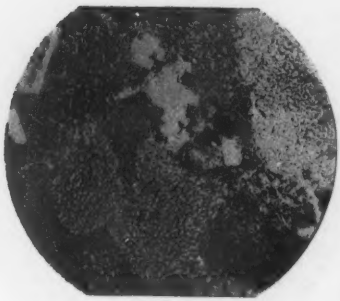


FIGURE 6

ORTHOCLASE PHENOCRYSTS IN RHYOLITE AND XENOCRYSTS IN CONTAMINATED ROCKS

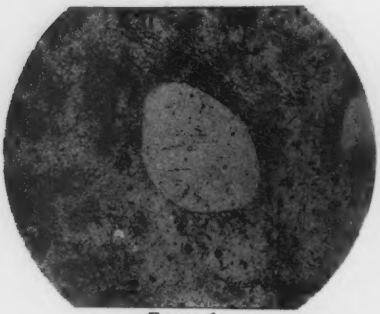
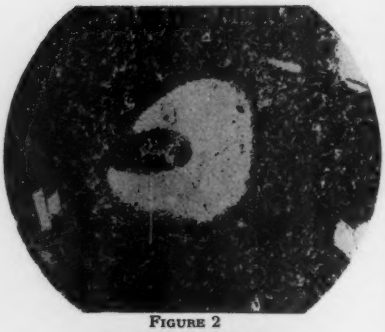


FIGURE 1



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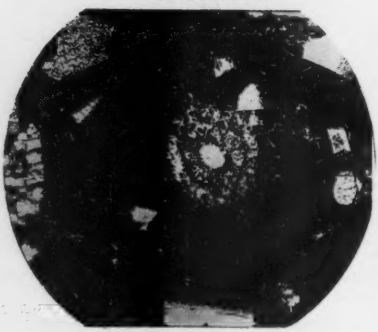

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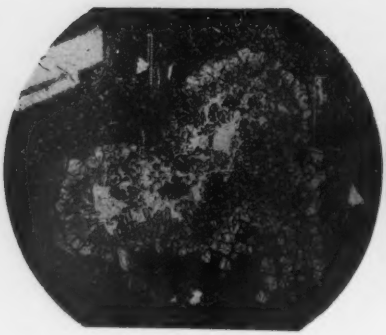

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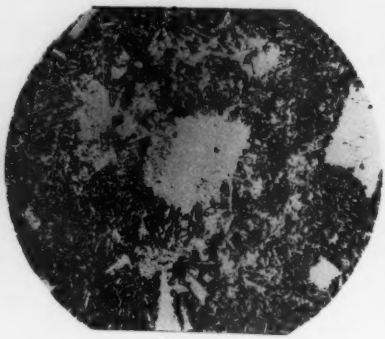


FIGURE 6

QUARTZ PHENOCRYSTS IN RHYOLITE AND XENOCRYSTS IN CONTAMINATED ROCKS

PLATE 6.—PHOTOMICROGRAPHS OF QUARTZ PHENOCRYSTS AND XENOCRYSTS IN CONTAMINATED ROCKS

Figure

1.—Rounded quartz phenocryst in granophyric rhyolite. (×19).

2.—Embayed quartz xenocryst in contaminated basalt enclosed in a thin zone of small augite crystals. An angular bytownite xenocryst is at the left. (×38).

3.—Large corroded quartz xenocryst in an indistinct rhyolitic streak in contaminated basalt enclosed in a zone of glass and small augite crystals. The mottled zone on the right is a "ghost" relic of orthoclase. (×17).

4.—Remnant of a small quartz xenocryst in contaminated basalt, 20 feet from contact with rhyolite. The compact rim of augite around the quartz is itself surrounded by a zone of glassy and cryptocrystalline material. (×17).

5.—"Ghost" relic of quartz in slightly contaminated basalt. Augite is the chief constituent, with material of low birefringence (tridymite?) in the central portion. (×34).

6.—Large xenocryst of quartz surrounded by wide zone of augite in slightly contaminated basalt Basaltic texture is seen at the edges of the photomicrograph outside the zone of augite. (×19)

PLATE 7.—PHOTOMICROGRAPHS OF PLAGIOCLASE PHENOCRYSTS IN CATARACT BASALT AND XENOCRYSTS IN CONTAMINATED ROCKS

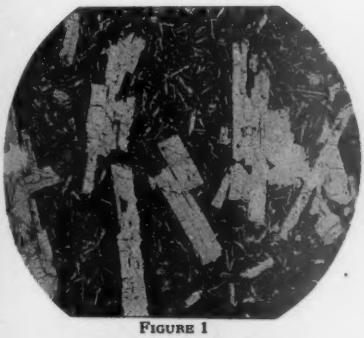
Figure

1, 2.—Bytownite phenocrysts in uncontaminated Cataract basalt. (×30).

3.—Zoned plagioclase in Cataract basalt. Outer portion is oscillatory-zoned bytownite. Inner portion (at extinction) is much more albitic than outer portion. This occasional zoning of crystals is characteristic of both the uncontaminated and contaminated Cataract basalt. (Crossed nicols, ×30).

4.—Euhedral and angular bytownite xenocrysts in rhyolitic rock. Optical characteristics show no marginal alteration or "albitization" in such xenocrysts. (×52).

5, 6.—Fracturing in bytownite xenocrysts. Fractures in such xenocrysts may be coated with dark material, but walls of fracture show no alteration. (×27).



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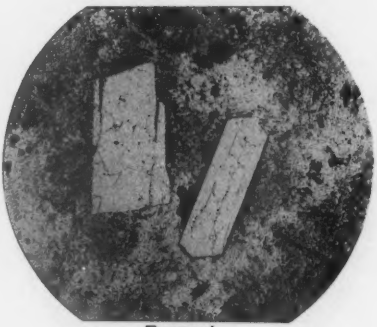

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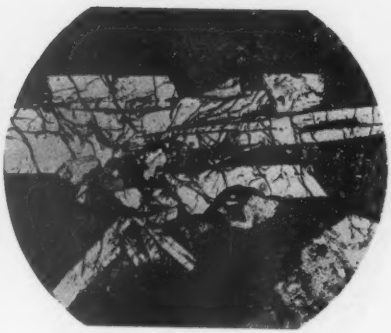
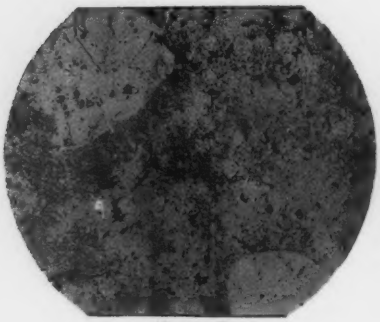


FIGURE 6

PLAGIOCLASE PHENOCRYSTS IN CATARACT BASALT AND XENOCRYSTS IN CONTAMINATED ROCKS



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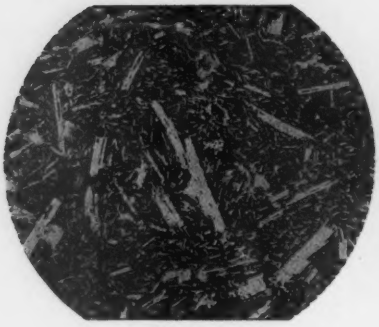

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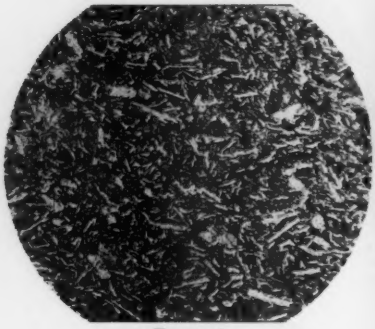

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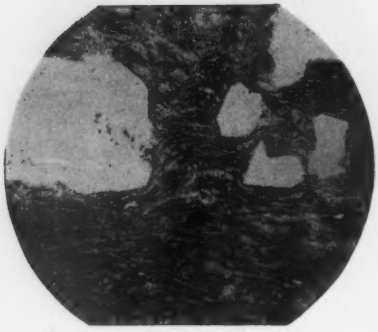

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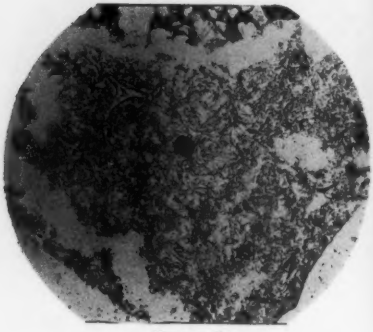

FIGURE 6

VOLCANIC ROCKS

PLATE 8.—PHOTOMICROGRAPHS OF VOLCANIC ROCKS

Figure

1.—Lodgepole rhyolite, uncontaminated phase. (×19).
2.—Cataract basalt, uncontaminated phase. (×23).
3.—Elkhorn basalt. (×23).
4.—Sheepeaters basalt (note lack of phenocrysts). (×23).
5.—Castle rhyolite, showing so-called "ignimbrite" texture. (×27).
6.—Meadow rhyolitic tuff. (×27).

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RHYOLITE-BASALT COMPLEX ON GARDINER RIVER, YELLOWSTONE PARK, WYOMING: A DISCUSSION

BY C. N. FENNER

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ABSTRACT

The immediately preceding paper in this issue of the Bulletin deals with a rhyolite-basalt complex on Gardiner River, Yellowstone Park, previously described by me (Fenner, 1938). Wilcox interprets relations in a very different manner, attributing the contact effects to the interaction of two simultaneous liquid lava flows. The present discussion adheres to the original view that the basalt is much older and that the rhyolite followed a canyon that had been erroded in the basalt. It is contended that all the contact effects can be explained logically on this basis, whereas Wilcox's view requires extraordinary and seemingly impossible behavior of the two magmas. Subsidiary phenomena likewise are out of accord with his explanation.

INTRODUCTION

The essential theme of Wilcox's paper is an interpretation of the rhyolite-basalt complex found in a rather small area on Gardiner River in Yellowstone Park. His view is much at variance with a description and interpretation that I gave in an earlier paper (Fenner, 1938).

The two rocks specifically in question are found along a rather narrow valley or gorge, having a rapid descent eastward (Pl. 1). In the river cliffs and in areas on the two sides a mingling of the rocks in great complexity is found. Farther back from the river on the north side unaltered rhyolite and basalt crop out; on the south side, in most places back from the river the surface is forested and soil-covered, but some basalt and rhyolite occur there also. It is evident that the mixed rocks are the result

of some sort of action of a body of rhyolite upon a body of basalt, but much of the main mass of rhyolite, which occupied the gorge where the river now flows, has been removed by erosion.

Unquestionably the rhyolite produced remarkable effects upon the basalt. On this we are agreed. The interpretation that I gave and to which I still adhere without reservation is that at some period prior to the appearance of the rhyolite thick masses of basalt had covered the area; that stream erosion had formed a valley here in close coincidence with the present gorge; and that a later flow of rhyolite followed the valley, producing contact effects and modifying the form of the valley to some degree by attacking and carrying away parts of the basalt in a manner indicated by the dikes, veins, and threads of rhyolite that penetrate the basalt. Later the gorge was re-excavated by removal of most of the softer rhyolite. By the action of the molten rhyolite upon the basalt great numbers of xenoliths were set free, and even large masses of undisturbed basalt have been soaked and modified in composition in the direction of rhyolite.

The rhyolitic magma was evidently highly gaseous, as is shown by the abundance of tridymite-lined vesicles and of tridymite in its groundmass. I expressed the view that the basalt was penetrated by tenuous rhyolitic solutions, liquid and gaseous, which in many places formed distinct veins cutting the basalt and elsewhere soaked large masses of rock, substituting rhyolitic for the original basaltic material in a manner analogous to that which has been recognized in the "granitization" of plutonic rocks.

The conception of Wilcox is that the rhyolite and basalt were essentially contemporaneous flows, that the basalt was molten when the liquid rhyolite came into contact with it, and that the rhyolite, because of its lower temperature, caused solidification of the basalt. As basis for his belief that the rhyolite could accomplish this he cites Larsen's conclusion that the temperatures of most crystallizing rhyolitic magmas are probably not far from 600° to 700°C., while those of most crystallizing basaltic magmas are probably close to 800° to 900°. For reasons that will be developed later I strongly doubt whether this gives sufficient theoretical basis for the observed phenomena and believe that in other respects the hypothesis encounters difficulties of the first magnitude.

The present trench of the river represents approximately the original form of the main mass of rhyolite, now removed. (Wilcox apparently agrees with this view.) It was therefore roughly a truncated pyramid lying on its side, pointing upstream, and encased in walls of basalt which were nearly vertical in most places. It is important to keep these relations in mind.

At the upper entrance to the gorge the immediately adjacent banks of basalt rise only 10–15 feet above the river, but the river descends rapidly eastward (Pl. 1), and at the lower end of the gorge the walls are 40–45 feet high on the left bank, a little lower on the right. The base of the rhyolite descends at about the same inclination as the river. The nearly vertical cliffs on the left (north) side of the river are especially to be noted. In places they are a complex mixture, but essentially they are basaltic masses which have been penetrated by rhyolite or on which remnants of rhyolite are plastered as a veneer. During the season of 1934, when most of my

work here was done, the river was abnormally low, and I had an exceptional opportunity to examine the character of these cliffs.

IMPOSSIBILITY OF SIMULTANEOUS LIQUIDITY AS SHOWN BY FORM OF CONTACTS

In proceeding to a more direct discussion of Wilcox's views I find a difficulty of this sort: that he expresses some of his fundamental ideas in such general terms that careful and repeated reading of his paper has not enabled me to visualize satisfactorily what he supposes conditions to have been at given stages of events. His idea that two fluid magmas came together is definite, but beyond that there is insufficient detailed description of supposed conditions that resulted. One must draw inferences as to his views from what is implied rather than stated. In important particulars it seems impossible to correlate such implications with what is seen on the ground, and one is left to wonder whether Wilcox himself formed a sufficiently definite picture.

The name Cataract basalt is given to the body with which the rhyolite reacted, and he supposes that after a longer or shorter period thereafter the "Sheepeaters basalt" appeared. To the rhyolite is given the name Lodgepole rhyolite. The idea seems to prevail in his paper that the meeting of the Cataract basalt and Lodgepole rhyolite was sufficiently violent to produce swirls and whorls of disintegrating basalt in rhyolite and the dragging out of one liquid by another in turbulent movement. I do not think that even if the two were simultaneously fluid their contact would have been of this character. The idea seems to imply some sort of restricted channel where the two flows met, and there is nothing in the geological surroundings to suggest this.

There is a canyon here at present, and along its walls we find the mixed rocks; but these walls are not some foreign rock that confined the flows but are themselves the Cataract basalt, which he supposes to have been liquid at the time.

Exposures of the Cataract basalt are found in a few places on the south side of the river for some distance back, and in many places on the north side for considerable distances. On this side their highest exposures pass under a small cliff of basalt, which may represent a second gush of the same flow or may be a separate flow (his "Sheepeaters basalt"). At this contact the Cataract basalt appears to have an essentially flat top. It is not known how far the Cataract basalt extends under the Sheepeaters basalt, but in the regions adjacent on the north, northwest, and northeast are the far-extended expanses of the Swan Lake flats, and there is no apparent obstacle to indefinite extension in this direction. The geological map of the U. S. Geological Survey indicates several square miles of the immediately contiguous part of this flat area to be underlain by Neocene basalt. Close to the gorge sufficient outcrops of the Cataract basalt are actually visible to show its character. It is a dense, homogeneous mass of rock, having no resemblance to the clinkery, slaggy, ropy flows that often proceed from central volcanoes and develop "moving stone walls." In its compactness it resembles plateau or fissure basalts, like most of the basalts of the Yellowstone region. Such basalts, in various regions, form smooth flows extending many miles, and by repeated extrusion they build up piles hundreds or thousands of feet thick. Where they flow over dry desert sands or over the surfaces of preceding flows their lower contacts seldom show evidence of brecciation and little evidence of

chilling effects of any kind. Whatever the reason may be, they are inherently different from the brecciated or scoria-forming basalts. If these plateau basalts are but little affected by surfaces 700° or 800° colder than themselves, what reason is there to attribute such remarkable chilling effects as Wilcox supposes to contact with melted rhyolite only 100° or 200° cooler?

Plateau basalts, in most of their occurrences, have a sheetlike form with approximately flat surfaces, and the contact with underlying sheets may be so obscure as to be almost unrecognizable. Wilcox seems to have had a very fluid basalt in mind. I would agree with this, and I believe that the evidence of its various fairly wide-spread exposures indicates that it was accumulating as a quiet lake, with nearly level surface. There is nothing in its character to suggest a departure from the ordinary features of plateau basalts.

We are now faced with a major difficulty involved in Wilcox's ideas, perhaps the most important of the whole problem, and the suggestions made by him in attempted solution seem to me very strange. His fundamental hypothesis is that the main body of fluid rhyolite formed a thick mass, enclosed in high, nearly vertical walls of liquid basalt. The density of rhyolite magma may be considered, closely enough, to be 2.36, and of basalt, 2.76. The Gardiner River rhyolite is shown by the almost microscopic thinness of many of the seams that penetrated basalt to have possessed great fluidity. This is a fact of observation, independent of any hypothesis as to the manner in which the seams may have been formed or of any conception of high viscosity that the rhyolite ought to have had. The low density and the high fluidity of the rhyolite should be considered in connection with Wilcox's statement that

"It is . . . to be emphasized that a general trough shape of the contact might . . . result from the meeting of basaltic and rhyolitic lavas, since a lobe of the rhyolitic lava would tend to 'float' in the basaltic lava during movement because of its less density."[1]

Reasoning of this kind leaves the reader bewildered. What sort of picture does Wilcox have in mind? The lesser density and the high liquidity of the rhyolite would certainly make it spread *over* the basalt and would *prevent* its forming a lobe. I should say that the tendency to float is the very reason why a thick lobe of rhyolite could not have been formed.

A flow of basalt, much more widespread than the present canyon walls (possibly covering square miles, so far as we can judge), had appeared in the region. Its depth at the time that the rhyolite came in (following Wilcox) is unknown, as we do not know how much basalt underlies the rhyolite, but the basalt was continuing to accumulate and finally reached a considerably greater height (that is, the eventual height was that at which this basalt passes under the "Sheepeaters basalt," probably 70 or 80 feet above the lowest basalt-rhyolite contact in the gorge). From some source a flow of thinly fluid rhyolite of relatively low specific gravity is assumed to have entered the area simultaneously. The natural supposition would be that it would spread over the surface of the basalt. Can one picture a reasonable method by which a lobe or block of it was caused to sink into the basalt and to be held in this

[1] This is quoted from a preliminary copy of Wilcox's manuscript that I had the opportunity of reading. It was eliminated in this form from the later copy, but the same idea is expressed in various places in the manuscript and, indeed, is the crux of the whole argument, so that the following discussion applies without change.

FIGURE 1. LOOKING EAST DOWNSTREAM

FIGURE 2. UPPER END OF GORGE
Basaltic ridge that was submerged and attacked by rhyolite.

FIGURE 3. PRECIPITOUS WALLS, ESSENTIALLY BASALT
Walls contain considerable rhyolite in places. Gently inclined shelf at top is basalt overlain by rhyolite.

VIEWS ALONG GARDINER RIVER

ATTACK OF RHYOLITE UPON BASIC INCLUSIONS, VALLEY OF TEN THOUSAND SMOKES

Showing antithesis of Wilcox's fundamental postulate. Xenoliths in fresh rhyolite became softened and disintegrated and pulled out into bands.

position while both it and the basalt continued to accumulate? I do not believe it possible. It seems as if it would be a major problem under Wilcox's conception to devise a means by which the mass of rhyolite could have taken the form that it did, but in his discussion we find little hint that such a problem is recognized.

Special attention is directed to the topography (Pl. 1, fig. 3) and particularly to the bench above the cliffs. This comprehensive view shows important features. Under the interpretation of either Wilcox or myself the main mass of rhyolite occupied the gorge, its base sloping at about the same grade as the river bed slopes, but some of it rose high enough to cover the bench. Most of the rhyolite has been removed, but locally on the bench it may still have a thickness of possibly 10 feet. On the opposite bank of the river also a thickness of several feet in places is indicated (Pl. 1, fig. 1), and on the vertical walls of the gorge remnants are plastered. In general, on the bench (Pl. 1, fig. 3), the rhyolite is quite thin, and ridges, domes, small jagged spires, and other minor irregularities of the underlying basalt stick up through it. Taking into consideration that most of the rhyolite has been removed we see that this bench and the walls of the gorge have essentially the original form of the underlying basaltic masses, not greatly masked by the residues of rhyolite. If all the remnants of rhyolite were stripped off, the character of the topography would not be essentially changed. Practically, we may consider all the rocks in this view as basalt and regard their present surface as the surface with which the rhyolite, in its gradual accumulation, came into contact. Were these walls and this bench liquid or solid at the time? It seems impossible to imagine that they were liquid. They appear in all respects like natural erosion features sculptured out of solid basalt. As a normal topographic feature the gently sloping bench with its small irregularities appears to represent a first shallow valley (Pl. 1, fig. 1) in which the gorge was subsequently cut by rejuvenated erosion. Up-river from the cascade is the untrenched old surface; in the opposite direction the river plunges deeper and deeper toward the profound depths of the main Gardiner River canyon. All this is in accord with a natural topographic development.

According to Wilcox, however, the place where we see the gorge is where a lobe of liquid rhyolite was accumulating against walls of liquid basalt. After the walls had been built up to the height of the rim of the gorge the rhyolite and basalt suddenly stopped accumulating along a vertical contact, and the rhyolite spread out over a gently sloping surface of basalt, in which upward-projecting ridges and domes and other irregularities rose from the general surface in some unexplained manner. The spreading of the rhyolite cannot be attributed to a cessation of accumulation of basalt, for the basalt is exposed (off to the right of the figure) at a considerably higher level than there is any trace of rhyolite.

In this interpretation, any appearance of topographic adjustment of the basaltic surface to its surroundings is wholly fortuitous.

Let anyone study the photograph and try to imagine a set of conditions that would bring about the relations that follow from Wilcox's postulates. One should bear in mind the lesser density of the rhyolite and the insulating properties of any wall that might conceivably be formed between the two. A sufficiently strong barrier to resist breakdown would be impossible. At a depth of 40 feet the differential hydrostatic

pressure on the basaltic side of a barrier would be about 1000 pounds per square foot, and for the full height of 40 feet it would be 20,000 pounds per running foot of wall.

The conclusions may be summarized as follows: According to my view the basaltic walls of the gorge and the surfaces of the adjacent area are an old erosion topography that was covered rather recently by a flow of rhyolite and was disinterred later in essentially its original form. It therefore accords with the general topography of the region. Wilcox's hypothesis would look upon the sharply cut gorge, the adjacent bench, and the minor irregularities not as parts of an erosion surface but the surface of a body of liquid basalt built up to different heights under a covering of liquid rhyolite. Erosion had nothing to do with the modeling of these basaltic surfaces. Over both the basalt and the rhyolite was later spread his "Sheepeaters basalt." The landscape of the whole territory would then have been very unlike that of today. Presumably it would be a wide-spreading, featureless lava field. During the succeeding ages the topography of the adjacent territory was carved out. In this immediate area the "Sheepeaters basalt" and most of the rhyolite that covered the Cataract basalt were stripped off. Then it was found, as a strange coincidence, that this basaltic surface was in topographic conformity with its surroundings. The profoundly deep canyon of the Gardiner to the north, eroding headward in this direction, entered the area at just the right place and at just the right level (at the bottom of the gorge) to bring everything into harmony with what would have been effected by erosion.

EXAMPLES FROM VALLEY OF TEN THOUSAND SMOKES

It is now proposed to examine Wilcox's idea of the solidifying effect of molten rhyolite upon molten basalt from another standpoint, that of the actual effect of fresh rhyolite upon certain basic rocks as seen in the Valley of Ten Thousand Smokes in Alaska. The rocks acted upon there are basic andesites rather than basalts, but the effects expected would not be very different. Evidence will be given to show that, instead of basic liquid lava having been solidified, blocks of solid lava were softened and disintegrated.

In the accumulations of tuff and pumice that cover the floor of the Valley, certain specimens of banded pumice, scattered over many square miles and numbering thousands, are striking and significant. Light bands, rhyolitic and containing crystals of quartz and albitic feldspar, alternate with dark or nearly black bands with many phenocrysts of pyroxene and calcic feldspar. These pumices have evidently been derived from the numerous explosion vents which were the origin of the tuff flows. At most of these vents ejection continued until the supply of available material was exhausted, but in the large vent of Novarupta in the upper part of the Valley the phase of violent eruption was succeeded by another phase. After the formation of a high ring of ejecta around the vent a great dome of banded lava was forced up in a highly viscous state. Here are to be seen in suspended operation and in diagrammatic clearness the processes that produced the banded pumices.

The rhyolites of the dome are not pumiceous but glassy. They have lost their pumice-forming gases. Inclosed within them are many blocks of basic andesites. These are of several types, but all have the textures of surface flows. The most probable source of them is in piles of morainal material that are believed to have lain

on the valley floor at the foot of the volcanoes, where Novarupta broke out. Direct evidence on this matter is concealed by the very thick deposits of ash and pumice that cover all the surroundings, but the character of the blocks points to such an aggregation. Material that even looks like glacial mud is included. In any event, great quantities of andesitic blocks became involved in the new rhyolite and were rapidly attacked by it. One may trace (Pl. 2) perfectly the transitions from angular blocks through softened and deformed clots and gobs to long-drawn-out bands. The great mass of Novarupta dome contains innumerable examples.

The material of the dome represents cessation of activity before attack on xenoliths was completed. The dark bands still contain phenocrysts of the basic minerals. In the pumiceous specimens scattered over the valley floor likewise this is generally true. Here violent ejection interrupted the process of attack, but in some specimens things have gone so far that phenocrysts of pyroxenes and basic feldspars are pretty evenly distributed throughout the mass in association with phenocrysts of quartz and albitic feldspar. In the pumices ejected from the main crater of Katmai, where basic rocks from the crater walls supplied the xenoliths, an even distribution of discordant minerals is the characteristic feature, and in some a final solution of the xenocrysts occurred and a gray pumice, nearly homogeneous, resulted.

The effects shown in the Katmai region are the antithesis of the effects that Wilcox ascribes to the meeting of rhyolitic and basaltic magmas. Instead of liquid basic magmas having been solidified, solid basic rocks have been softened and disintegrated. If one should maintain, nevertheless, that contrary effects may sometimes be produced, the burden of proof seems to be to bring forward equally convincing examples.

Magmas should not be regarded as inert mixtures, such as are worked with in experimental laboratory melts. They are extremely complex chemical systems, formed of many nonvolatile and volatile constituents. During their ascent from great depths, changes of internal equilibrium lead to exothermal reactions, somewhat analogous to the development of the latent heat of crystallization. Some such source of heat is required to explain contact modifications and other phenomena associated with them.

TEXTURE AND STRUCTURE OF MIXED ROCKS

The mixed rocks are of two general types. In many places these types are closely associated, but the effects are fairly distinct. In the first, basaltic rocks are penetrated by definite dikes, veins, and threads of rhyolite. In the second, large and small masses of basalt of uniform appearance have been modified in composition in the direction of rhyolite.

A typical example of the first type is found on the south side of the river near the head of the first cataract. The river has here cut through a low ridge of basalt and its accompanying complex. In Figure 2 of Plate 1 the ridge lies in the trees on the right, but basaltic rocks come down to the water's edge. The bare expanse at the left of the figure is the complex. At the top of the bench here the rock abutting against the ridge is mostly rhyolite, but it contains many distinct xenoliths of basalt. Downward, the proportion of basalt increases irregularly, and at the water's edge, 25 or 30 feet from the top of the bench, there is almost nothing but basalt, but it is still pene-

trated by a few small veins of rhyolite. Likewise from left to right the amount of basalt increases by transition into the massive basalt of the ridge. In simplest terms, the ridge at the right is a uniform mass of slightly modified basalt whose left flank has been penetrated by a complex of irregularly ramifying and intersecting veins of rhyolite. This grades off to the left and above into nearly pure rhyolite. The veins range in width from a maximum of about a foot in the upper portions to minute dimensions at the bottom. The same relations continue along the river bank to the left for a considerable distance.

The logical explanation of what we see here seems to be that when the flow of rhyolite came into contact with the long-solidified basalt tenuous solutions penetrated inward and downward along incipient cracks, causing them to open up. Near the original contact surface many xenoliths were set free, but in depth the penetration of rhyolite gradually ceased. At an early stage of reaction considerable basalt may have been carried away by the rhyolite, but the stage that we now see represented is one of little or no movement.

Farther down the river a knob or column of basalt, probably 8–10 feet high, rises from the river bed, isolated from the adjacent high bank except for a large dike of very vesicular rhyolite that joins the column to the basaltic cliffs. The contact of rhyolite and basalt is approximately vertical, but in detail it is very irregular. The rhyolite corrodes the basalt at the contact and on the left side sends small stringers and a fairly large horizontal band or irregular protuberance into it. The main part of the rhyolite also contains corroded inclusions of basalt. On the right side of the dike the basaltic cliff is intricately veined with rhyolite. These veins are exposed for a considerable distance along the cliff and then die out in basalt. The main vertical dike of rhyolite, in continuing downward, finally breaks up into stringers in basalt at water level.

Can such relations be explained by the mingling of two fluid magmas? I think not. Many complexes of the same character are found in the area, and in all of them the veins of rhyolite in basalt have no resemblance to the flowing forms of a mixture of liquids but are definitely the filling of cracks in solid rock, irregularly corroded at contacts. They bear a close resemblance in many respects to the reticulating stockworks of ore deposits.

In places these vein injections have penetrated a good many feet into the basalt. Near the foot of the high cliff on the left bank of the river a cave at one place goes back into solid rock for perhaps 10 feet. The cliff face on the outside is rhyolite carrying occasional boulderlike inclusions of basalt, but this rhyolite is only a veneer, for after one passes the rhyolite arch at the entrance the rock within the cave is basalt seamed with rhyolite. The cave does not run back far enough to show the limit of penetration of these veins, but on the bench at the top of the cliff just above the cave the same complex is exposed, extending about 40 feet back from the brink. Here the basalt was exposed to attack on two sides, from rhyolite in the gorge and from rhyolite above it on the bench.

Many other examples of this type of penetration might be described, but though they differ endlessly in detail their major features are in most cases much the same. In some of the veins the rhyolite is nearly pure; elsewhere disintegration products of

basalt contaminate it greatly. Then the rhyolite becomes dark, and casual inspection might give the impression that two liquids had mixed, but microscopic study of thin sections shows that this is not true (Pl. 3, figs. 3, 4). Bits of basalt, varying in size from microscopically large masses to minute specks are scattered through the rhyolite. Separate crystals of basic feldspar, pyroxene, and olivine, much corroded, are associated with such magmatically incongruous minerals as quartz, tridymite, and alkalic feldspar. Nests or vugs of tridymite crystals are common, both in the larger rhyolite masses and in the veins that cut basalt. The veins vary commonly from a width of 1–2 feet down to almost invisible seams. (Certain large rhyolite dikes will be described separately.) Their range is from the top of the cliffs to the lowest rocks in the gorge—that is, about 45 feet—, but in the lowest exposures they are definitely petering out. I do not know of any feature in which these veins differ from what might be expected of offshoots from the main body of rhyolite working into cracks in the solid basalt that encloses it and corroding their walls. Many features are similar to certain phenomena of "granitized" rocks.

In the exposure shown in Figure 1 of Plate 3 rather different features are found. The mass of rock outlined is a much modified basalt, 100 feet or more long, jutting out as a shelf over rhyolite that is nearly pure but contains some boulders of basalt. At the top of the cliff the same modified basalt is clearly exposed over an area 20 or 30 feet broad. A few feet farther back normal basalt crops out.

This mass I interpret as part of a cliff that was undercut by the flow of rhyolite. The contact between rhyolite below and modified basalt above is irregular on a small scale, but regular on a large scale. It is well defined and knife-sharp. For some distance below the contact the rhyolite contains many angular xenoliths. Portions of basalt at the contact are nearly surrounded by veins of rhyolite and appear ready to drop.

The main mass is basalt, most of it still normal looking but other parts greatly modified. The whole mass, from top to bottom, is seamed with rhyolite veins that ramify through it in the most intricate manner. A specimen taken at the contact was analyzed. In appearance it has been greatly altered and is now of a uniform, medium gray color and of fine, even grain, reminding one of the xenolithic "ghosts" found in masses of granitized rock, but there is no doubt that it belongs with the basalts. A distinguishing characteristic of the modified basalts is that they may shade gradually into more normal varieties or may be of uniform appearance, but the contacts of all varieties with rhyolite are sharp. Analysis showed that the specimen had been impregnated with rhyolitic material to the extent of about 68.5 per cent. The great alteration of this specimen is attributed to its position immediately over a large body of rhyolite and probably not far from another large mass of rhyolite that filled the adjacent gorge. It was in a specially favorable position for impregnation by rhyolitic solutions, which may have been largely gaseous.

So far as I can see, there is no possible way in which the mass of thoroughly impregnated and greatly modified rock outlined in the photograph could have retained its position overlying rhyolite if it was liquid or mushy at the same time that the rhyolite was liquid. Even its most altered parts are of greater density than the rhyolite, as witnessed by the xenoliths dropping from the contact and frozen in place, as

described above. If it be contended that this great mass had been almost instantaneously solidified by contact with rhyolite (which itself seems an impossibility), then the very phenomena of veining and impregnation of basalt by rhyolite that Wilcox's hypothesis has been devised to explain are left without an explanation.

A question may arise in the reader's mind, nevertheless, whether at least some of the complexes of the area could not have been formed as breccias where molten rhyolite and molten basalt came into contact in the way that Wilcox supposes. A difficulty in answering this is that Wilcox himself does not give a consistent picture of the "modus operandi" of such action. He seems to suppose that at times contact resulted in a uniform mixture of magmas and at other times breccias were formed. These ideas seem to be antagonistic. Whatever picture Wilcox may have had, the idea of simultaneous liquidity of two magmas, uniform mixing, and contemporaneous brecciation does violence to ordinary laws of physics and mechanics.

Comparison has been made in this discussion of some of the features of the Gardiner River complexes with those of stockworks of ore and with granitization phenomena. A stockwork is generally understood to mean a mass of rock intersected by many small veins of ore and gangue, sometimes nearly parallel, sometimes crossing and interlacing in all directions. The fragments of country rock between veins or surrounded by veins may be highly mineralized throughout, or an outer shell of replaced rock may surround kernels that are relatively unchanged. In granitization a similar network of veins may, in favorable examples, intersect the country rock, shading off on one side of such a complex into an igneous-looking mass and on the other dying away in unaltered country rock; but in granitization the newly introduced minerals are silicates instead of ore and gangue. Their resemblance lies in the fact that both stockworks and granitization have resulted from the invasion by extraneous solutions of masses of solid rock, not only along cracks, but by penetration and soaking along the most minute interstices between grains. Thus the masses have been gradually metasomatized, and the "islands" of original rock, while retaining approximately their outlines as units, have been gradually worked over in composition. It is this sort of action that I believe is responsible for the basalt-rhyolite complexes in the Gardiner River area. During prolonged contact of a gas-charged rhyolitic magma the solid basalts have been invaded and metasomatized. In any of these forms of replacements—stocksworks, granitization, or the Gardiner River complexes—individual specimens may appear like breccias; in fact, they all are breccias if the term be understood to mean simply a rock broken into fragments in any manner. So far as I can grasp Wilcox's idea, however, he has had something quite different in mind, apparently a breccia produced by sudden chilling and shattering of liquid basalt at surfaces of contact with cooler rhyolite. Against this idea we have to consider not only what we see in a hand specimen (which might have been caused in a variety of ways) but the form and relations of a complex as a whole to its surroundings, and in preceding pages I have tried to describe several examples of these relations in such a way that the reader may visualize them and see the impossibility of their having been formed in such manner as Wilcox implies.

Further consideration of "granitization" of these rocks will be given a little later in connection with the second type of modification that has affected them.

FIGURE 1. MASS OF VEINED AND METASOMATIZED BASALT
Undercut by rhyolite

FIGURE 2. PSEUDO DIKES OF RHYOLITE
Cutting basalt cliffs

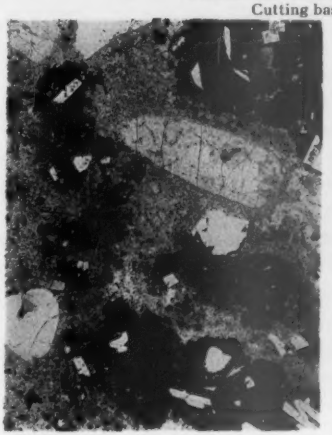

FIGURE 3. BASALT DISINTEGRATING IN RHYOLITE
Plain light. x 20.

FIGURE 4. BASALT DISINTEGRATING IN RHYOLITE
Plain light. x 20.

CONTACTS BETWEEN RHYOLITE AND BASALT

At several places the basaltic cliffs on the north side of the river are cut by dikes or pseudo dikes larger than the veins previously described. They have a width of 2–3 up to 8–10 feet. They are a peculiar feature but they look like the filling of vertical clefts, somewhat modified by rhyolitic corrosion (Pl. 3, fig. 2). (At the right edge of the photograph the top of a third dike is shown, which passes down out of the picture. At A is a heterogeneous mixture of rhyolite with xenoliths of basalt, lying on the upper bench. The contact of the rhyolite of the dikes with their basaltic walls is very sharp.)

Wilcox's explanation of these dikes does not offer anything that one can grasp. Does he suppose that a solid wall of basalt was first formed, and that this suddenly cracked and let rhyolite rush in? Under such conditions, or anything similar, the lake of liquid basalt behind the wall would be tapped, and the basalt, being heavier, would rush out. The greatest confusion would result.

A specimen of uniformly soaked basalt taken from the top of the wall between the two right-hand dikes, just under the capping of rhyolite, was analyzed and contained about 13 per cent of rhyolitic material. A large, angular, sharply bounded xenolith of basalt enclosed in the rhyolite near the bottom of the middle dike had 30 per cent rhyolite.

These much altered but uniform xenoliths of basalt illustrate the second type of modification, as distinguished from definite veins. On a larger scale this modification appears as a uniform impregnation of cliffs or other large masses of basalt with rhyolite. In these the action has not proceeded as far as in the smaller examples described, which represent material surrounded by rhyolite or otherwise in a specially favorable position. Xenoliths in rhyolite veins are likely to be altered in the more intense manner, and innumerable specimens are available. They are readily recognizable by their lighter color but have sharp contacts with rhyolite. The larger masses are not so easily recognizable as different from normal basalts. An analysis may be required to reveal a different composition, or quartz and alkalic feldspar may show up in microscopic study.

Wilcox finds difficulty in understanding how phenocrysts (I would say metacrysts) of quartz and alkalic feldspar could be introduced into solid basalt and believes that such associations imply actual mixture of basaltic and rhyolitic liquids. Here, as in other portions of his paper, I find discussion of the matter hampered by the fact that he does not present a clear picture of how he supposes the action to have taken place, especially as two incongruous ideas seem to be involved. In places he seems to imply that, when the basaltic and rhyolitic magmas met, the basaltic liquid was frozen almost immediately, and thus a breccia or a solid wall of separation was formed; elsewhere it is implied that a uniform mixture of the two liquids on a large scale was effected. As a matter of fact both these processes would have had to occur to bring about the results that we see if the hypothesis of two contemporaneous magmas is maintained, but nevertheless they seem incompatible.

The term "granitization" has been widely but rather loosely used in the literature. The British geologists, Arthur Holmes and Doris Reynolds, have termed the processes implied by it "mobilization." Their conception is that various constituents of a magma may become "mobile" or transferable into other rocks, while at the same time

the constituents of the other rocks yield to the invading forces and pass on. Thus a series of changes is effected. Dunn has called rocks resulting from such processes diabrochites—soaked rocks. A somewhat similar idea was urged as long ago as 30 or 35 years by Sederholm, Lacroix, F. D. Adams, and a few other leaders. In later years the idea has been more fully developed, and many authors have cited examples. One might infer from Wilcox's treatment of the subject that in my advocacy of the replacement of basaltic minerals by rhyolitic I had attempted to introduce a new principle in petrology. The arguments that he brings to bear against it are of such sweeping character that they would apply not only to the Gardiner River rocks but in equal degree to all the examples of granitization that many competent geologists have studied and described; indeed, they would apply nearly as well to metasomatic processes in general. Is that the stand that he wishes to take?

As an example of his attitude on such processes I quote from his manuscript:

> The postulate that vapors were driven into the solid basalt from the rhyolite lava requires that a pressure gradient must have been present through the basalt. The process of modification must have ceased at some time, yet where are the basaltic constituents (the reaction products) that were being removed when the pressure and temperature gradient degenerated? Admittedly some should have been trapped during the dying stages of the activities of these vapors, but none are present, as has already been noted. . . . A critical inconsistency becomes apparent. If the vapors escaped with their load to an outside destination through the solid basalt wall . . . , we must also consider the progress of the vapors that were penetrating the larger basalt xenoliths from all sides. The reaction products would then have been carried to the center of the xenoliths and, having no further escape would have been deposited there. When we look for these reaction products . . . , we find no trace of them but only a graduation into unaltered basalt."

Instead of trying to formulate by abstract reasoning what should have taken place under his view of conditions, he might well have given attention to the fairly copious literature on granitization and considered carefully what does take place. If his reasoning were correct the objections he expresses would apply equally well to the inclusions in plutonic rocks, but the bleached "ghosts" of uniformly metasomatized xenoliths that occur in great abundance in granitized zones are not imaginary objects, nor are the large crystals or groups of crystals of alkalic feldspar that have been developed in the midst of quartzites or hornblendites to be explained by a mixing of magmas. His idea of what should or must occur is not in accord with the conception that geologists who have worked in these fields would apply. Their view is that the invasion of emanations into the solid walls advances along a fairly definite front, with temperature and chemical action at a given point progressively rising as the emanations advance. Thus the wave of advancing emanations drives ahead of it more and more of the original constituents of the rock. These emanations move along the easiest avenues of escape and may reach the surface. Under plutonic conditions they may eventually be so dispersed as not to be recognizable, but what may be their ultimate destiny in some cases has been discussed in a series of papers in the Journal of the Geological Society of London that have attracted much attention. Professor Holmes and especially Miss Reynolds (Reynolds, 1944) have attributed the formation of certain mafic rocks of "igneous" aspect to the addition of constituents thus mobilized and driven out of granitized rocks.

Returning to the Gardiner River rocks specifically, there are other vital objections to Wilcox's view that uniform mixtures of basaltic and rhyolitic minerals must have

been formed by mixtures of the corresponding magmas. The rocks under immediate consideration are not those which are penetrated by networks of definite rhyolitic veins, but those in which impregnation is nearly or quite uniform. In some of these a vague and almost imperceptible dappling or streakiness may be seen on close examination, but in many instances even this is absent. It is hard to imagine magmas mixing so perfectly that uniformity should be so nearly complete, especially if one magma tends to chill and brecciate the other. Their sharpness of contact (lack of gradation) with rhyolite is much against it. Still more important is their microscopic texture. Basalts in general have a well-recognized, distinctive texture, peculiar to themselves. If a basaltic magma, free or nearly free from groundmass minerals, as Wilcox supposes, were uniformly mixed with a considerable amount of rhyolitic magma the result would be an andesitic magma, and nothing in our experience allows us to suppose that the resulting crystallized rock would have a typical basaltic texture and typical basaltic minerals. Nevertheless, such textures and such minerals are characteristic of these modified rocks until replacement is far advanced. A typical feature of such rocks is the persistence of groundmass pyroxene, olivine, and feldspar, arranged in basaltic texture, but with individual outlines corroded as they gradually yield to quartz and tridymite and alkalic feldspar. Frequently, sharply bounded, angular xenoliths, enclosed in rhyolite, show an advanced stage of replacement in an outer shell, but progressively lessening and giving place to essentially unchanged basalt in the interior. Such phenomena are generally looked upon as evidences of metasomatic replacement. They mean that before the rhyolitic material was introduced the rock was a solid basalt. I do not believe there is any escape from that conclusion. The magmatic incompatability of large amounts of labradorite, olivine, iron ore, and pyroxene with quartz and tridymite is unquestionable. Certainly this would not be the result if basaltic and rhyolitic *magmas* were mixed and allowed to crystallize. Such a mixture of molten liquids would have the composition of an andesite and would crystallize as an andesite. The excess silica supplied by the rhyolite would have combined with the orthosilicate molecules of potential olivine and probably with part of the iron oxides of the ore, and the resulting magnesium-iron metasilicate would give rise to a pyroxene of composition different from that of the original lime-magnesium-iron metasilicate of the basalt. It is plain that what has actually happened with these contaminated rocks is that a basaltic magma crystallized in normal fashion, with appropriate texture and minerals, and that into this solid rock rhyolitic material was introduced, replacing part of the original basalt. (Wilcox's Plate 6 shows several good examples of the preservation of basaltic texture in contaminated rocks.)

Thin sections of uncontaminated basalts were studied particularly to see whether any trace of free silica minerals—quartz or tridymite—could be discovered, and none was found.

The preservation of basaltic textures is a common feature of these uniformly contaminated rocks, but it naturally loses sharpness as the amount of rhyolitic material introduced becomes great, but even in the specimen of uniformly medium-gray rock from the contact shown in Figure 1 of Plate 3, in which analysis shows 68.5 per cent of rhyolitic material, olivines are present.

A good example of the uniform impregnation of large masses of basalt is a specimen taken from a cliff face about 10 feet from the nearest rhyolite exposure. Analysis shows an addition of 12.7 per cent rhyolite. It is uncertain to what distance rhyolitic material has soaked into these masses of rock, for they generally show no megascopic difference from normal basalt.

Thus, homogeneous mixing of basaltic and rhyolitic magmas should have produced andesites instead of the rocks of basaltic textures and basaltic minerals with salic minerals dispersed through them.

There may be hesitation in accepting the idea of a process akin to "granitization" to explain effects produced by surface lavas, but I believe that the importance of this area lies precisely in the fact that it demonstrates in clear and unmistakable fashion that under favorable circumstances such processes do occur.

SIGNIFICANCE OF LODGEPOLE OBSIDIAN

In Wilcox's classification of the rocks of the area the rhyolite that reacted with the basalt is called the Lodgepole rhyolite. His Lodgepole obsidian is regarded as a phase of the same flow, and he lays much stress upon the fact that, as he believes, this flow took the glassy form only where it came into contact with the older cold Meadow rhyolitic tuff, and that where it is in contact with the then molten Cataract basalt it is felsitic in texture.

A little consideration of the obsidians and felsites elsewhere in the Park should have demonstrated that such a simple explanation of the formation of obsidian is not valid. The towering Obsidian Cliff is glass from top to bottom, but only the base could have been chilled by contact with cold rock. On the other hand, the common felsites of the Park do not have a glassy base. In the case of the Lodgepole obsidian specifically Wilcox seems to have failed to visualize the consequences that would have resulted from his own hypothesis. This obsidian may rest on tuff (though this is doubtful) but if it is contemporaneous with the Cataract basalt, as he supposes, the upper part of its mass must have accumulated in contact with molten basalt.

The main exposure is a few feet or tens of feet outside the precipitous cliffs of the gorge, but the basaltic flow could not have stopped there abruptly. Near-by exposures of Cataract basalt overtop the obsidian by many feet, and there must have been exactly the same sort of contact between the basalt and the obsidian (except at its base) that there was between the basalt and the felsitic rhyolite in the gorge. Moreover, should not the material underlying the obsidian have been heated by passage of earlier portions of the basaltic and rhyolitic flows? In fact, if the glassy phase of the rhyolite has any significance in the problem, it means that the Cataract basalt was *not* molten at the time.

In the latter part of Wilcox's paper and in the bibliography is given a long list of examples of contemporaneous or nearly contemporaneous ejections or intrusions of salic and mafic magmas, as found in the literature. These show the possibility of such occurrences (which I would not question and, in fact, I have cited such phenomena in other papers as a problem for the advocates of crystal fractionation to solve), but they prove nothing regarding contemporaneity of the Gardiner River rocks. Moreover, in respect to some of the examples that he gives there is strong reason to

doubt contemporaneity. In others, the arguments used by their authors to explain the introduction of salic minerals into mafic rocks are of the sort that has been the subject of severe criticism by Professor H. H. Read (1944). Discussing a paper by Reynolds on the Newry igneous complex Read stated his opinion that

> "The production of oligoclase porphyroblast diabrochite [impregnated or soaked rock] by the growth of closely packed porphyroblasts of oligoclase was a process well known from innumerable migmatite zones. This was the process of felspathization that had been demonstrated on geological evidence dozens of times. . . . The refusal to admit the validity of the plain geological evidence had often led to absurd proposals, such as, for example, the mechanical forcing in of felspars as such into xenoliths and marginal country rocks. It was true of course, that once one admitted the validity of felspathization, one was logically compelled to consider the possibility of a non-magmatic origin of many so-called igneous rocks, and it might be this sequel that caused the reluctance of the magmatists to modify their views."

In recent British petrological literature there has been manifest a reaction against the tendency to explain the widest variety of igneous or apparently igneous phenomena by fractional crystallization. The latter theory seemed at one time to be carrying everything before it. The reaction is a healthy development, if it does not go so far as to eliminate magmatic processes entirely from the field in explaining plutonic rocks, as seems to be threatened in some instances.

SUMMARY

In the present paper discussion of Wilcox's views has been directed to the major features of the problem. In respect to many of the minor features also I find myself in disagreement. In some cases this is a matter of fact finding; in others, of interpretation of facts. However, conclusions really depend upon the major features, and to go into details would prolong the discussion to a degree that is hardly warranted.

A brief summary of conclusions is as follows:

(1) There is no evidence that there was any sort of restricted channel through which two liquid lavas were forced to flow, or anything that would bring about turbulent mixing. On the contrary, when the flow of basalt occurred it apparently spread over a wide area, and if the rhyolite was liquid at the same time there is no apparent reason why it should have been concentrated where we now find it.

(2) A fluid rhyolite of relatively low density could not have been forced down as a thick mass into the heavy basalt and held there; or, if this seemingly impossible condition were brought about, we are left without an explanation as to why the rhyolite, in its continued accumulation, spread as a gently inclined layer over the rim of the gorge.

(3) The present form of the basaltic surface, as revealed by the later stripping off of rhyolite, is in accord with normal topographic development of the adjacent region and with headward erosion of the main Gardiner canyon. A fortuitous accord of a basaltic surface congealing from a liquid would be a remarkable coincidence.

(4) The idea, fundamental to Wilcox's argument, that a rhyolitic magma could chill and solidify a basaltic magma upon contact is not supported by any evidence. Results observed in the Katmai region, where inclusions of basic rocks involved in rhyolite became softened and disintegrated, is the antithesis of this action.

(5) The dikes and veins of rhyolite that penetrate basalt, taken individually or in

the groups that form complexes, are of the character that would result from the entrance of rhyolite into cracks or fissures in solid rocks, with concomitant corrosion, soaking, and partial replacement of the solid structure. The phenomena are very similar to stockworks of ore or zones of granitization, and the replacements are in accord with the general principles of metasomatism. Wilcox's arguments against this interpretation seem to have been developed without consideration of the many descriptive and explanatory discussions in petrologic literature.

(6) Large and small basaltic masses modified by uniform permeation of rhyolitic material, still show characteristic basaltic textures and minerals, together with quartz, tridymite, and alkalic feldspar dispersed through them and corroding the original mafic minerals. This association is magmatically incompatible. Such textures and minerals would not have been developed from a uniform mixture of basaltic and rhyolitic magmas.

(7) Wilcox stresses the felsitic character of the rhyolite, where it is in contact with basalt, and the occurrence of glassy obsidian where such contact is lacking. He attributes the difference of character to a heated condition of the basalt when the rhyolite flow took place. All such significance is lost when it is realized that if the two flows had been contemporaneous the basalt would not have stopped as a high cliff but would have continued the few feet necessary to bring it into contact with the obsidian.

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RECONSIDERATION OF THE MORRISON FORMATION IN THE TYPE AREA, JEFFERSON COUNTY, COLORADO

BY W. A. WALDSCHMIDT AND L. W. LeROY

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ABSTRACT

The section of the Morrison formation recommended as the "type" has been selected because the entire Morrison is exposed along the West Alameda Parkway road cut, 2 miles north from Morrison, Colorado, and because the locality is readily accessible.

The Morrison formation at the recommended type section is a distinct cartographic unit which can be subdivided into six parts primarily on the basis of lithologic characteristics of the sedimentary rocks.

Fossils in this section of the Morrison formation are confined to three horizons. Dinosaur remains occur in limited number in basal sandstones of the gray shale and sandstone unit. Probable fresh-water sponge spicules are plentiful throughout an interval of 25 feet in the central part of the gray clay and limestone unit. A few spicules, however, are also present from 2 to 6 feet above the dinosaur-bearing sandstone. *Aclistochara*, the occurrence of which has not been mentioned previously in literature on the Morrison formation in Colorado, is relatively abundant in the upper 30 feet of the gray and red shale unit. This would place the highest occurrence of *Aclistochara* 63 feet above the base of the Morrison formation.

The results of copper nitrate staining of samples from the calcareous rocks of the Morrison formation indicate that the greatest amount of calcite is present in the gray clay and limestone unit. In all probability the distribution and intensity of copper nitrate stains could be used effectively in connection with petrographic studies to identify several strata within the formation.

INTRODUCTION

Early published descriptions of the type Morrison formation are relatively brief and generalized; therefore, the authors present detailed descriptions of the many lithologic components of the formation at its newly proposed type section and also describe several units which can be identified on the basis of paleontological evidence, stain tests, and physical characteristics.

The authors recommend that the exposure along the north side of the West Alameda Parkway road cut (SE¼ Sec. 23, T.4S., R.70W., Morrison quadrangle, 7½ minute series, Jefferson County, Colorado) (Pl. 1, fig. 1), 2 miles north of Morrison, Colorado, be accepted as the revised type section of this formation. This recommendation is made because the descriptions of the type section of the Morrison, at or near Morrison, Colorado, based on incomplete exposures, have been inadequate. Furthermore, at the recommended type section the Morrison formation is exposed in its entirety and it can be found readily (Fig. 1).

The Morrison formation, as herein described, is the cartographic unit between the Dakota sandstone as defined originally by Eldridge (1896, p. 62) and the strata assigned to the probable Sundance by Lee (1920, p. 184) which have in part been assigned to the Ralston formation by LeRoy (1944). The Morrison as defined in this paper is considered Jurassic.

REVIEW OF INVESTIGATIONS

First mention of sediments, which undoubtedly included those assigned to the Morrison formation by Eldridge, was that of Hayden (1869, p. 136). In his survey of the Turkey Creek area (2 miles south from Morrison) he commented,

> "On the western side of this ridge [referring to the Dakota hogback] we see the projecting edges of the sandstone capping the ridge, and underneath the variegated marls and sandstones, with some of the brick-red beds."

Marvin (1874, p. 93) described strata, outcropping at Bear Creek, later considered as Morrison, as "variegated red and green shales, arenaceous and argillaceous". He also mentioned a "white sandstone" which is the equivalent of the basal sandstone of the type section proposed in this paper.

The Morrison formation was first named, described, and defined by Eldridge (1896, p. 60–62), although Cross (1894, p. 2) was the first to use the name Morrison for equivalent beds in his work on the Pikes Peak quadrangle in central Colorado.

Eldridge's description of the Morrison at the type section is as follows:

> "It is a formation of fresh-water marls of an average thickness of about 200 feet. Its upper limit is sharply defined by the Dakota sandstone, while the brown and pink sandstone closing the Trias [basal Morrison sandstone of the present paper] as clearly marks its lower limit. To this formation has been assigned the name "Morrison" from the town near which it is typically developed."
>
> "The marls are green, drab, or gray, and carry in the lower two-thirds numerous lenticular bodies of limestone of a characteristic drab color and a texture compact and even throughout .... The clays of the lower two-thirds of the Jura are remarkable for their reptilian remains, and from the predominating form have been designated 'Atlantosaurus Clay'."

In connection with the position of the Morrison-Dakota contact there has been some controversy (Lee, 1920). Eldridge (1896, p. 63) comments that the base of

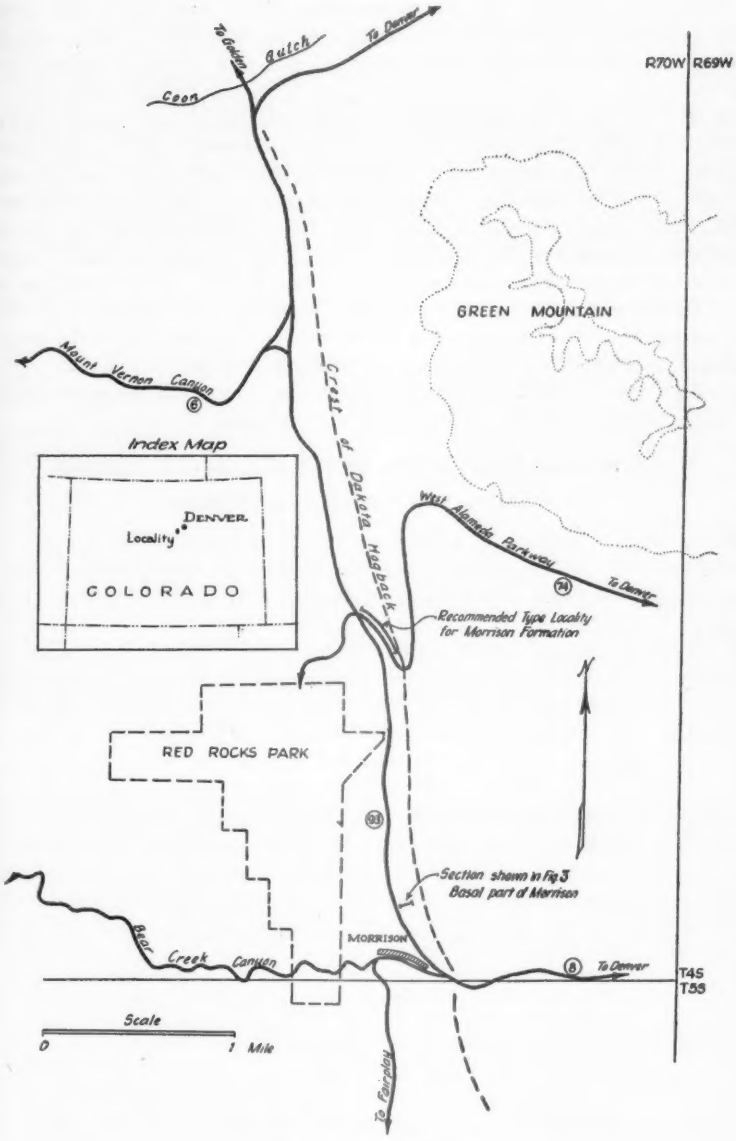

FIGURE 1.—*Index map*

Showing location of recommended type locality of the Morrison formation, Jefferson County, Colorado.

the Dakota is marked by a distinct conglomerate which "varies from a thin, almost imperceptible layer to 30 feet in thickness, and is composed of well rounded, smooth, in some cases almost glazed, pebbles from the size of a pea to a diameter of 1 inch."

Lee (1920, p. 184), in a rather confusing paper, redefined the Morrison at its type locality. It appears that in his reconsideration he erroneously interpreted Eldridge's Morrison-Dakota boundary to be somewhat stratigraphically higher than Eldridge had originally indicated. Consequently, Lee proposed that the contact be lowered to Eldridge's original boundary. Later Lee (1927, p. 28) re-examined the Dakota and Morrison formations near Morrison, Colorado, and presented a more complete discussion of the lithic units and their relationships. In this work he designated the lowest Dakota unit as the "Lower Sandstone" (10–40 feet) and described it as "massive, conglomeratic". This boundary corresponds to that proposed in the present paper and is believed by the writers to be as Eldridge originally defined it.

Since Lee's work in 1927 there has been no further information published pertaining to the Morrison formation of the type area in Colorado.

RECOMMENDED TYPE SECTION OF THE MORRISON FORMATION

STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS

The type section is along the north side of the West Alameda Parkway road cut, SE¼ Sec. 23, T.4S., R.70W., Morrison quadrangle, 7½ minute series, Jefferson County, Colorado, 2 miles north of the town of Morrison. It is 277 feet thick at this locality.

As defined in this paper, the Morrison lies with apparent disconformity below the conglomeratic phase of the Dakota as it was originally defined by Eldridge (1896, p. 60–62) and Lee (1927, p. 28) and overlies disconformably the red, sandy shales assigned by various workers to the Lykins formation of Triassic age. At some localities the basal sandstone of the Morrison is in juxtaposition with strata that have been correlated by Lee (1920, p. 184) with the Sundance (Jurassic) and which have been recently included within the Ralston formation by LeRoy (1944).

The upper and lower lithic boundaries of the Morrison are distinct in the recommended type section as well as in adjacent areas. Outcrops of the Morrison are sporadically exposed along the west slope of the Dakota hogback north and south from the type section. The formation at the type section dips 25° E. and strikes N. 15° W.

LITHOLOGY

The Morrison of the recommended type section can be subdivided into six lithologic units which in descending order are as follows:

(1) Sandstone and Shale Unit: This unit has a thickness of 76 feet 6 inches. Variegated sandy shales (maroon most prevalent) constitute about 30 per cent of the interval. The sandstones are essentially gray to buff, massive, friable, and medium-grained. At the base of the unit is a 15-foot bed of thinly stratified sandstone (Pl. 1, fig. 3) which contains, in the lower 2 feet, several gray sandy shale layers none of

FIGURE 1. COMPLETE MORRISON FORMATION
Exposed along the West Alameda Parkway road cut, facing Mt. Vernon Creek, 2 miles north of Morrison, Jefferson Co., Colorado.

FIGURE 2. CONTACT BETWEEN MORRISON AND DAKOTA FORMATIONS
See columnar section (Fig. 2).

FIGURE 3. BASAL SANDSTONE OF SANDSTONE AND SHALE UNIT
76 feet 6 inches below base of Dakota.
See columnar section (Fig. 2).

FIGURE 4. BASE OF CALCAREOUS SANDSTONE BEARING DINOSAUR REMAINS
See columnar section (Fig. 2).

FIGURE 5. INTERBEDDED GRAY CLAY AND LIMESTONE
See columnar section (Fig. 2).

RECOMMENDED TYPE SECTION OF MORRISON FORMATION

FIGURE 1. LOWER PART OF MORRISON FORMATION
Facing Mt. Vernon Creek, State Highway 93, 0.2 mile north of Morrison, Jefferson Co., Colorado.
Basal sandstone in lower left and *Aclistochara* beds in center and upper right.

FIGURE 2. *Aclistochara* FROM BASAL PART OF MORRISON FORMATION
Same area as Figure 1. LD-88, 89. × 60.

LOWER MORRISON FORMATION

which exceeds a thickness of 3 inches. The upper 2 feet of the 15-foot bed contains many small limonitic concretions which weather out to give a pitted surface.

(2) RED SHALE UNIT: This unit has a thickness of 36 feet 9 inches, consists of massive red to maroon sandy shales, and contains several beds of buff, medium-grained, calcareous sandstone layers whose maximum thickness is 1 foot. This unit is the most highly colored interval within the formation.

(3) GRAY SHALE AND SANDSTONE UNIT: Sediments of this stratigraphic unit have a thickness of 51 feet 6 inches and consist of gray clays and several 1-foot beds of gray calcareous sandstone. In the lowest part are a few thin beds of gray limestone which are either sandy, sandy-dolomitic, or chalcedonic. The basal sandstone of this unit, 8 feet thick, is gray to buff, calcareous, coarse-grained, frequently conglomeratic, and contains dinosaur remains (Pl. 1, fig. 4). The extremely uneven lower surface of this sandstone indicates a probable formational hiatus; however, it cannot be evaluated at this writing.

(4) GRAY CLAY AND LIMESTONE UNIT: This unit, an interbedded series of gray clays and gray, lithographic, fresh-water limestones, is 49 feet 9 inches thick. Many of the clay layers are highly indurated yet contain very little or no calcium carbonate cement. The induration is probably a result of compression and internal rearrangement of the original constituents, elimination of free water, or mechanical packing accompanied by dehydration of aluminum silicates and colloids. The clay-limestone ratio is 4:1. The limestones rarely exceed a thickness of $1\frac{1}{2}$ feet and average 6 inches. Two beds of gray, medium-grained, calcareous sandstone, each 2 feet thick, occur in the upper third of the unit. The base of this unit is arbitrarily placed at the highest observed occurrence of *Aclistochara*. Most thin sections of the limestone layers examined contain numerous blade-shaped objects which have been tentatively identified as fresh-water sponge spicules.

(5) GRAY AND RED SHALE UNIT: This unit, representing the lower 55 feet of the Morrison, consists of gray and red silty shales. The red silty shales are prevalent in the lower half. Three thin limestones occur in the upper half of the unit. Species of *Aclistochara* are irregularly distributed through this interval; however, they are abundant and well developed in the upper third. *Aclistochara latitruncata* Peck, *A. obovata* Peck, *A. lata* Peck, and *Chara voluta* Peck have been identified. A few poorly preserved fresh-water gastropods and ostracods were also noted. This unit is also exposed in an outcrop 0.2 mile north from Morrison.

(6) BASAL SANDSTONE UNIT: The base of the Morrison is marked by 7 feet of buff, massive, locally cross-bedded, coarse- to medium-grained, calcareous sandstone, conglomeratic in its lower part. North and south from the type locality this sandstone thickens to at least 30 feet.

DETAILED SECTIONS MEASURED AT TYPE LOCALITY

The following detailed descriptions are of the sedimentary rocks constituting the recommended type section of the Morrison formation exposed along the north side of the West Alameda Parkway road cut, in the $SE\frac{1}{4}$ Sec. 23, T.4S., R.70 W., Jefferson County, Colorado (Fig. 2).

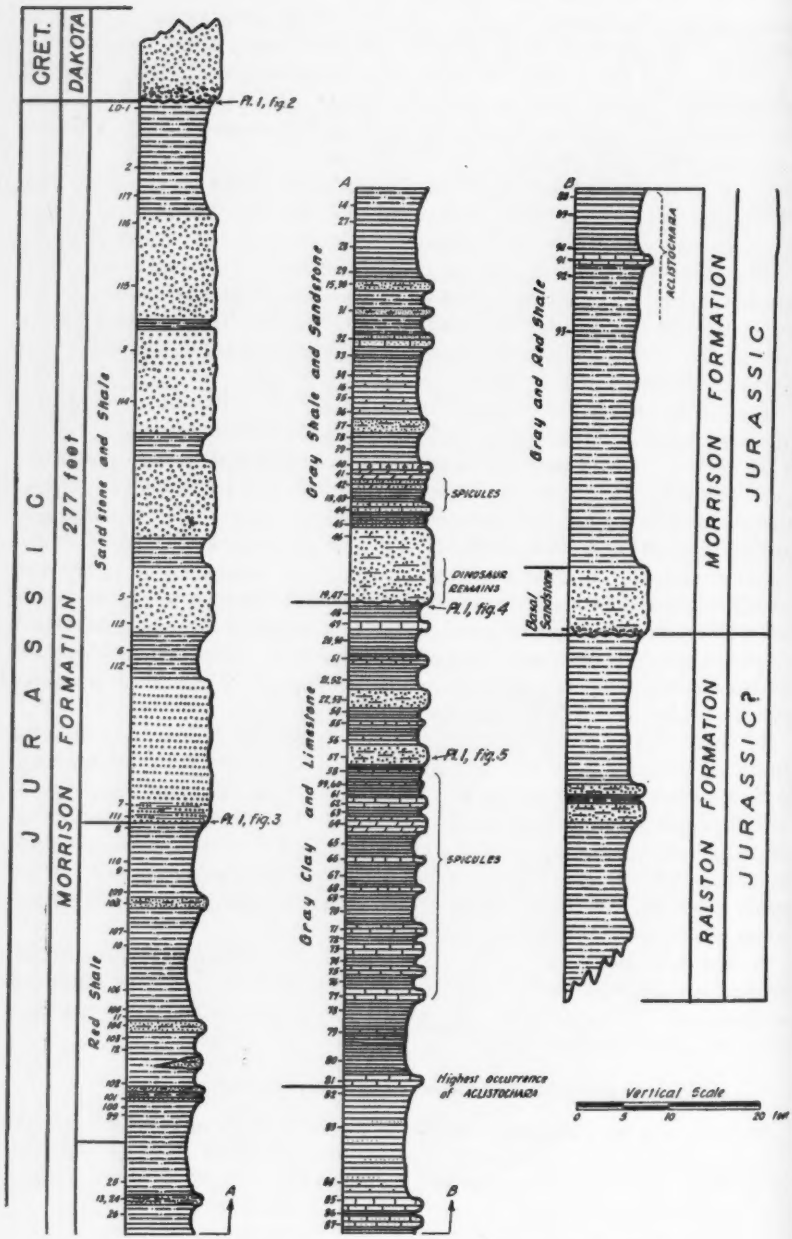

FIGURE 2.—*Columnar section of Morrison formation at recommended type locality*

Alameda Parkway Road cut, 2 miles north from Morrison, Jefferson County, Colorado. Numbers refer to samples in collection series LD. on file at the Colorado School of Mines, Golden, Colorado.

DAKOTA

MORRISON

		Feet	Inches
	SANDSTONE AND SHALE UNIT		
Shale:	Slightly sandy, maroon and light lavender intermixed. Lavender shale predominant in top 6 inches, somewhat platy; contains few poorly preserved bladelike plant impressions. LD-1*	6	
Shale:	Sandy, greenish-gray, faint lavender, pale maroon. LD-2, 117	6	
Sandstone:	Yellow to buff, massive, medium- to coarse-grained, friable, with ½-inch streaks of chocolate-brown, hard, ferruginous sandstone at top. Surface coated brownish red from overlying shale. LD-115, 116	11	
Shale:	Greenish gray to gray, sandy.	1	
Sandstone:	Light buff to pink, massive, slightly porous, friable, medium- to fine-grained, pink discoloration on surface. Several large concretions, maximum 1 foot in diameter. LD-3, 114	11	
Shale:	Sandy, gray and maroon intermixed.	3	
Sandstone:	Buff, massive, medium-grained, friable. Small ferruginous concretions in lower 2 feet.	8	
Shale:	Gray, maroon, yellow intermixed; sandy, moderately bedded.	3	
Sandstone:	Pale buff, massive, fine- to medium-grained, loosely compacted. Small ferruginous concretions in upper 1½ feet. LD-5, 113	7	
Shale:	Sandy, greenish gray to gray. LD-6	2	
Shale:	Dark maroon, sandy, poorly bedded, blocky fracture, some greenish-gray streaks. LD-112	3	
Sandstone:	Grayish-white to light pink, fine to medium, compact, small specks of limonite; well bedded with individual beds ranging from 6 inches to 1 foot. Small limonitic concretions in top 2 feet. Lower 4 feet exceptionally well bedded; contains numerous layers of light gray, slightly gritty, hard, compact shale. LD-7, 111.	15	6
	Total	76	6
	RED SHALE UNIT		
Shale:	Light gray to faint red, hard, compact, sandy. Sand grains fine to coarse. LD-8	1	
Shale:	Gray to maroon, sandy, blocky fracture, white, speckled, sand grains medium. LD-9, 110	4	
Shale:	Maroon, sandy, blocky fracture. LD-109	3	
Sandstone:	Buff, medium-grained, calcareous, massive. LD-108	1	6
Shale:	Variegated light gray, pale yellow, light lavender; color distribution uneven; compact, massive, very irregular fracture.	1	
Shale:	Maroon, sandy, blocky fracture.	2	
Shale:	Gray, slightly sandy, massive, compact, irregular fracture, slightly speckled, slightly calcareous. LD-10, 106, 107	7	
Shale:	Pale greenish-gray to pinkish, gritty, very hard, compact, massive, uneven to subconchoidal fracture. LD-11, 105	3	
Sandstone:	Yellow to brown, calcareous, limonitic, massive, fine-to medium-grained, compact. LD-104	1	
Shale:	Light gray to pale buff, mottled, sandy, compact, irregular fracture. Pale maroon in lower foot. Two feet of medium-grained friable sandstone lense in middle part. LD-12, 102, 103	6	
Sandstone:	Light buff, massive, slightly calcareous, medium grained, friable.	1	

*Numbers refer to sample series LD on file in the Department of Geology at the Colorado School of Mines, Golden, Colorado,

		Feet	Inches
Shale:	Maroon, massive, uneven fracture, sandy.	0	9
Sandstone:	Yellowish-brown streaked with gray; calcareous, fine grained, massive, compact. LD-101	0	6
Shale:	Maroon, compact, poorly bedded. LD-100	2	
Shale:	Greenish-gray, sandy.	1	
Shale:	Maroon to greenish-gray; compact, irregular fracture. LD-99	2	
	Total	36	9

GRAY SHALE AND SANDSTONE UNIT

Shale:	Light gray, massive, sandy, compact, irregular fracture, slightly mottled. LD-25	4	
Shale:	Gray, sandy, slightly calcareous, massive, hard, compact, uneven fracture. LD-13, 24	1	
Clay:	Gray, slightly sandy, hard. LD-26	1	
Clay:	Gray, massive, hard, locally slightly calcareous. LD-14, 27, 28, 29	12	
Sandstone:	Gray, fine- to medium-grained, argillaceous, massive, compact, slightly dolomitic. LD-15, 30	1	
Shale:	Gray, sandy, imperfectly laminated.	2	
Sandstone:	Gray, massive, argillaceous, calcareous. Contains sandy clay inclusions. LD-31	0	6
Clay:	Gray, sandy, compact, uneven fracture.	2	
Limestone:	Gray, sandy, impure, brittle, irregular fracture. LD-32	1	6
Clay:	Gray, compact, hard, irregular fracture. LD-33	2	
Clay:	Bluish-gray, soft, locally slightly calcareous. LD-16, 34, 35, 36	5	6
Sandstone:	Gray, argillaceous, slightly calcareous. LD-37	1	6
Clay:	Gray, soft, slightly calcareous. LD-38, 39	3	
Limestone:	Light to dark brown, crystalline, dolomitic, numerous veinlets filled with chalcedony. LD-40	0	6
Dolomite:	Gray to light brown, limy, fine-textured, speckled with limonite. Weathered surfaces pitted. LD-41	1	
Dolomite:	Light tan, sandy, slightly argillaceous, few spicules. LD-42	1	
Clay:	Gray, massive, uneven fracture, hard. LD-18, 43	1	6
Limestone:	Yellowish-brown to gray, sandy. Sand grains fine to medium. Numerous spicules. LD-44	1	
Limestone:	Gray, sandy, dolomitic, some limonitic streaks, few feldspar grains. LD-45	1	6
Sandstone:	Light gray, coarse, variable grain size, angular to subangular, highly calcareous. Dinosaur remains in basal 2 feet. LD-19	8	
	Total	51	6

GRAY CLAY AND LIMESTONE UNIT

Clay:	Gray, massive, soft, uneven fracture. LD-48	2	
Limestone:	Light tan to gray, lithographic. LD-49	0	6
Clay:	Gray, massive, poorly bedded. LD-20, 50	3	
Limestone:	Light gray with dark gray patches, conchoidal fracture, lithographic. LD-51	1	
Clay:	Gray, massive. Several 8-inch concretions. LD-21, 52	2	
Sandstone:	Gray, medium to coarse, calcareous. LD-22, 53	2	
Clay:	Gray, massive, soft. LD-54	1	6

		Feet	Inches
Limestone:	Light gray, lithographic, some pyrite. LD-55	0	6
Clay:	Gray, silty, slightly calcareous	1	
Clay:	Gray, massive, slightly calcareous. LD-56	1	
Sandstone:	Light gray to brown, massive, dolomitic, fine-grained, compact, quartzitic appearance. LD-57	2	
Clay:	Gray, massive.	0	6
Limestone:	Gray to pale yellowish-brown, spicules common, pyrite rare. LD-58	0	3
Clay:	Gray, massive.	1	
Limestone:	Yellowish-gray, lithographic, some pyrite. LD-59, 60	0	3
Clay:	Gray, massive. LD-61	1	
Limestone:	Light gray, lithographic, conchoidal fracture, sandy in lower 6 inches, numerous spicules. LD-62	1	
Clay:	Gray, massive. LD-63	0	6
Limestone:	Mottled yellow to light gray, dolomitic, some anhydrite, brecciated appearance, spicules common. LD-64	2	
Clay:	Gray, massive, poorly bedded. LD-65	3	
Limestone:	Light gray, lithographic, some anhydrite, few quartz grains, numerous spicules. LD-66	0	6
Clay:	Gray, massive, poorly bedded. LD-67	3	
Limestone:	Light gray to pale tan, fine even texture, spicules common but unevenly distributed. LD-68	0	3
Clay:	Gray, massive, slightly calcareous. LD-69, 70	3	
Limestone:	Light gray, fine texture, spicules common. LD-71	1	
Clay:	Gray, massive, slightly calcareous. LD-72	1	
Limestone:	Medium to light gray, yellow specks, very fine texture, spicules common. LD-73	1	
Clay:	Gray, massive. LD-74	1	
Limestone:	Light gray, massive, fine even texture, spicules common. LD-75	1	
Clay:	Gray, massive. LD-76	2	
Limestone:	Light gray, massive, very fine crystalline, spicules common. Spicules rare below this point. LD-77	1	
Clay:	Gray, massive. Slightly calcareous in lower 4 feet. Ostracod sp. LD-78, 79, 80	8	
Limestone:	Pale yellowish gray, yellow-brown specks, massive, very fine crystalline, some anhydrite; few *Aclistochara* (highest recorded occurrence); spicules rare. LD-81	1	
	Total	49	9

GRAY AND RED SHALE UNIT

Clay:	Gray, massive, locally sandy. LD-82, 83, 84	12	
Limestone:	Light gray to tan with slight pale-green tint, very fine texture, few calcite-filled fractures, spicules rare. LD-85	1	
Clay:	Gray, massive. LD-86	0	6
Limestone:	Light gray with faint greenish cast, slightly sandy, fine even texture, spicules rare. LD-87	1	
Clay:	Gray, massive, poorly bedded. *Aclistochara* in lower 5 feet. LD-88, 89, 90	8	
Limestone:	Light gray, massive, fine even texture, spicules rare. LD-91	1	
Clay:	Gray, massive, sandy to silty. *Aclistochara* in top foot. LD-92, 93	8	
Shale:	Red to gray, color unevenly distributed, mottled appearance.	24	
	Total	55	6

		Feet	Inches
	BASAL SANDSTONE UNIT		
Sandstone:	Buff, massive medium- to coarse-grained, calcareous, locally cross-bedded, red on surface from overlying shale. LD-23	7	
	Total	7	0
	Total thickness of formation	277	

DETAILED SECTIONS OF BASAL PART OF MORRISON FORMATION

The following detailed descriptions are of the basal part of the Morrison formation (Fig. 3; Pl. 2, fig. 1), 0.2 mile north from Morrison, on the west side of State Highway 93, Jefferson County, Colorado. (Supplemental to recommended type section.)

		Feet	Inches
Clay:	Gray, massive. LD-119	2	
Limestone:	Gray, lithographic, weathers light brown. *Aclistochara* rare. LD-118	2	
Clay:	Gray, massive; *Aclistochara* scarce. Few thin shelled mollusk fragments. LD-120	0	6
Clay:	Gray, massive, indurated. *Aclistochara* scarce. LD-121	1	
Clay:	Gray, massive; *Aclistochara* scarce. LD-122	0	6
Clay:	Gray, indurated. *Aclistochara* common. LD-123	0	6
Clay:	Gray, sandy, slightly calcareous. Orange and yellow chert grains common. *Aclistochara* rare. Few smooth ostracods. LD-124, 125, 126	4	
Shale:	Purple, massive, blocky fracture.	0	4
Shale:	Gray, silty, massive. LD-127	2	6
Clay:	Gray with local faint maroon streaks. *Aclistochara* present. LD-128, 129	5	
Limestone:	Gray, clayey to silty. LD-130	2	
Clay:	Gray, massive.	1	
Clay:	Gray to faint maroon, slightly indurated, blocky fracture. Top of lower maroon clays. LD-131	3	
Shale:	Red to gray, massive, slightly calcareous. LD-132	1	
Shale:	Faint maroon, silty, massive. LD-133	3	
Shale and impure limestone:	Irregularly interbedded, gray to maroon. LD-134	2	
Shale:	Maroon, massive; *Aclistochara* and ostracods rare. LD-135, 136, 137	12	
Sandstone:	Buff, massive, medium-grained, calcareous	2	
Shale:	Maroon, slightly silty	6	
Sandstone:	Buff to gray, stained red on exterior, massive, locally cross-bedded, friable, jointed.	31	
	Total thickness	79	4

ACLISTOCHARA

Considerable importance is attached to the presence of Gyrogonites of *Aclistochara* in the lower part of the Morrison formation. In the West Alameda Parkway section *Aclistochara* (Pl. 2, fig. 2) is confined to 58 feet of the section directly above the basal sandstone. In the section just north from Morrison the genus is distributed throughout 50 feet of the formation just above the basal sandstone unit. In other Morrison sections in adjacent areas *Aclistochara* is restricted to this portion of the Morrison;

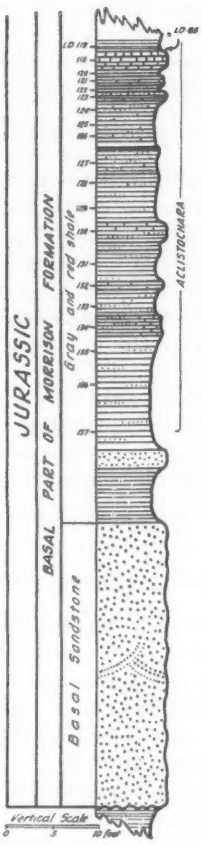

FIGURE 3.—*Columnar section of the lower part of the Morrison formation (supplemental to the recommended type section)*

State Highway 93, 0.2 mile north from Morrison, Jefferson County, Colorado.

however, a few specimens have been observed locally in the shaly phases of the underlying Ralston sediments. It is of special interest to note that several species of *Aclistochara* recorded by Peck (1937, p. 83–90) from the Morrison of Wyoming were also identified in the sections discussed in this paper. The forms described by Peck seemingly occur in approximately the same part of the Morrison as those herein noted.

In future work on the Morrison sediments, *Aclistochara* and related forms should be investigated carefully as such forms may prove to have stratigraphic significance.

STAIN TESTS

Samples from 28 strata of the Morrison formation, between the base of the red shale unit and the lowest limestone bed in the gray and red shale unit, were stained with copper nitrate for the purpose of identifying calcite in sandstones and for differentiating calcite from dolomite in the highly calcareous beds. A polished flat surface of each specimen was etched with dilute hydrochloric acid (1–20), boiled for 10 minutes in 1 molar copper nitrate solution, washed in running water, and immersed in ammonium hydroxide for 5 minutes to darken and fix the green copper nitrate stain.

Enough variation of distribution and intensity of the green stain was observed so that it is believed that several of the limestone beds in the Morrison type section can be identified by staining tests alone. However, the stain tests may prove more useful in identifying strata within the Morrison formation if they are used with other features of the rocks which can be observed in thin sections or hand specimens. (*See* Table 1.)

PETROGRAPHIC DESCRIPTIONS

Thin sections were prepared from the same samples which were used for stain tests. Detailed descriptions of all the thin sections would be too voluminous for the purpose of this paper; therefore, only the most important features are here briefly summarized and are also given in Table 1. A limited number of photomicrographs were made to illustrate some effects of staining, characteristics of sand grains, and features observed in thin sections (Pls. 3, 4).

Aclistochara was not observed in any of the thin sections prepared from samples in which *Aclistochara* could be seen on weathered surfaces. Sponge spicules were readily observed in thin sections but were not discernible in the hand specimens from which the thin sections were prepared. One stratum, 14 feet above the base of the gray shale and sandstone unit, contains chalcedony with fine radiating structure (Pl. 3, fig. 1). Chalcedony of this type was not found in other strata within the Morrison formation. Dolomite, as a cementing material, is relatively abundant in the sandstones within the gray shale and sandstone unit and is more abundant within this unit than in the underlying gray shale and limestone unit. Small amounts of anhydrite were observed in only three thin sections prepared from samples of the strata in the gray shale and limestone unit. Other features in these sections are given in Table 1.

TABLE 1.—*Stain tests and petrographic features*

Sample No.	*Classification*	*Stain effect*	*Petrographic features*
LD-24	Indurated sandy shale	None	Even distribution of fine sand grains.
LD-30	Argillaceous, dolomitic sandstone	None	Uniform distribution of dolomite. Occasional fine clay inclusions.
LD-31	Argillaceous, dolomitic sandstone	None	Similar to LD-30.
LD-40*	Cherty limestone	Good; shows many unstained areas of chert and clay	Chalcedony with fine radiating structure as filling of veinlets and cavities.
LD-41	Dolomite	Faint, uneven greenish stain	Extremely fine crystalline texture. Small, poorly developed dolomite rhombs throughout mass.
LD-42	Argillaceous dolomite	Faint greenish	About equal amounts argillaceous material and dolomite.
LD-44	Argillaceous sandy dolomite	Moderate, uneven pale greenish	Brecciated appearance; moderate amount of sand; few spicules and argillaceous parts only.
LD-45	Dolomitic sandstone	Pale brownish green	Wide spacing of variable-sized grains in fine-grained matrix. Occasional minute dolomite crystal.
LD-46	Calcareous sandstone	Moderate, uneven	Grains variable size, 0.36 mm. maximum diameter. Wide spacing of calcareous cement. Many chert grains.
LD-47	Calcareous sandstone	Moderate. Confined to calcareous cement	Similar to LD-46.
LD-49	Lithographic limestone	Good; even	Extremely fine texture. Dolomite crystals and sand grains rare.
LD-51	Lithographic limestone	Good; slightly mottled	Extremely fine texture. Abundant fractures filled with calcite and chalcedony.
LD-55*	Lithographic limestone	Excellent; first section with a very distinct green stain; faintly mottled	Little dolomite unevenly distributed.
LD-57	Calcareous sandstone	Good; confined to cement	Similar to LD-46 and LD-47 but with slightly more dolomite as cement.
LD-58	Lithographic limestone	Good; slightly uneven	Very fine texture; small scattered areas of dolomite; spicules.
LD-60*	Lithographic limestone	Excellent green with only faint mottling	Extremely fine texture.
LD-62	Lithographic limestone	Good; slightly mottled	Very fine texture; few minute sand grains. Occasional spicule.
LD-64*	Dolomitic limestone	Good; mottled	Brecciated appearance; numerous spicules; irregular small patches of dolomite. Small amount anhydrite.
LD-66	Lithographic limestone	Moderate; slightly uneven	Fine texture; numerous spicules; very little anhydrite; occasional patch of dolomite.

TABLE 1.—*Concluded*

Sample No.	*Classification*	*Stain effect*	*Petrographic features*
LD-68	Limestone	Moderate; faint mottling	Fine texture; moderate amount of spicules; dolomitic areas relatively fine crystalline.
LD-71	Limestone	Good; mottled; few unstained spots	Fine texture; numerous spicules. Many areas dolomite or magnesium rich calcite. Wavy extinction in calcite.
LD-73*	Limestone	Excellent; faint mottling	Very fine texture, few spicules, few small areas dolomite, little anhydrite.
LD-75	Limestone	Excellent; faint mottling	Very fine texture; numerous spicules.
LD-77*	Limestone	Excellent; mottled	Very fine texture ground-mass with 30–40% of coarser dolomite and calcite grains 0.06 to 0.12 mm. in diameter. Few spicules.
LD-81*	Limestone	Excellent with brown unstained spots	Aggregates of dolomite grains which have hollow appearance because of limonitic-filled fractures radiating from center.
LD-85	Limestone	Moderate; slightly mottled	Fine-grained groundmass in which dolomite appears to have replaced the limestone along small irregular fractures.
LD-87*	Limestone	Excellent; even except for small patches	Fine texture; slightly sandy; few spicules.
LD-91*	Limestone	Excellent except for few brown unstained areas of dolomite	Fine texture with rounded to rhombic crystals of dolomite. Similar to LD-81.

* Strata which might be identified on basis of stain tests and thin-section examinations. Sample numbers correspond to those on the columnar section (Fig. 2).

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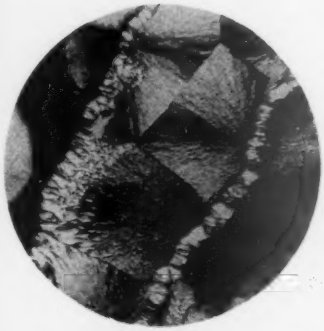
EXPLANATION OF PLATES 3–4

PLATE 3.—PHOTOMICROGRAPHS OF THIN SECTIONS

Figure

1. Chalcedony in crystalline limestone, 6 feet above sandstone containing dinosaur remains. Sample No. LD-40. ×60. × nicols.
2. Calcareous sandstone from upper part of stratum containing dinosaur remains. Illustrates rounded form of grains and relatively wide separation of grains by calcareous cement. Sample No. LD-46. ×60. × nicols.
3. Calcite fracture filling in lithographic limestone. This type of filling is common in many of the lithographic limestones in the Morrison formation, but the width of the fractures and the size of the calcite grains vary. The texture of other lithographic limestones in the Morrison formation is similar to that shown on the right and left sides of the illustration. Sample No. LD-51. ×60. × nicols.
4. Probable fresh-water sponge spicules in limestone. Forked spicules such as shown near the center of the illustration are rare. Small straight spicules which are evident around the forked form are abundant in most of the limestones of the gray clay and limestone unit. Sample No. LD-68. ×60, × nicols.
5. Dolomite crystals in thin limestone, fifty-five feet stratigraphically above basal Morrison sandstone. Sample No. LD-81. ×60. × nicols.
6. Dolomite crystals in thin limestone, thirty-two feet stratigraphically above basal Morrison sandstone. Sample No. LD-91. ×60. × nicols.



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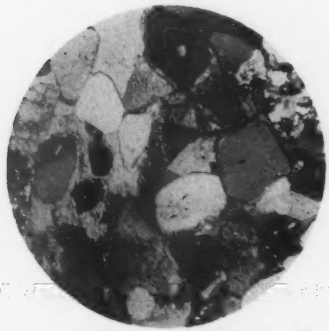

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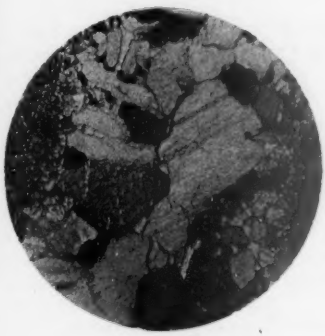

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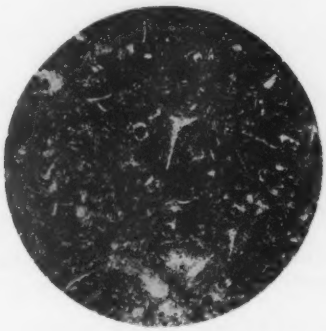

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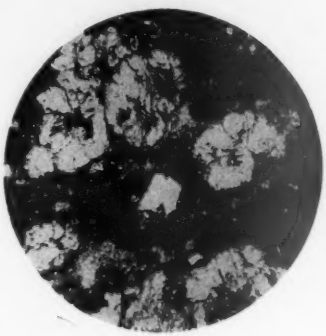

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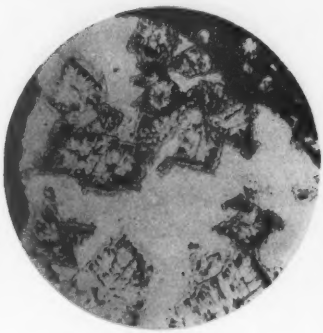
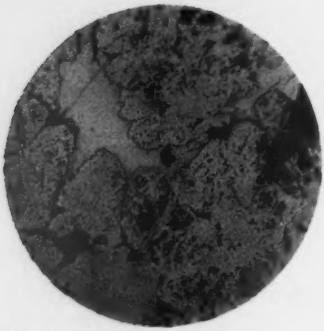


FIGURE 6

PHOTOMICROGRAPHS OF THIN SECTIONS



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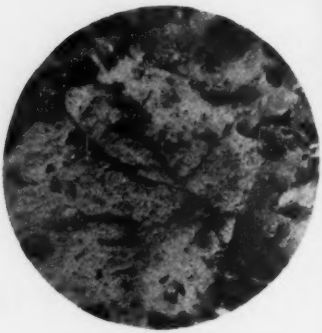

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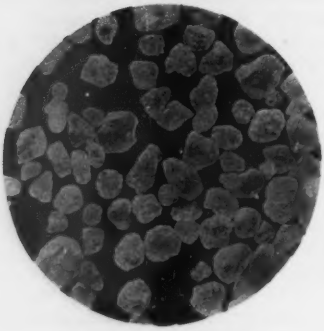

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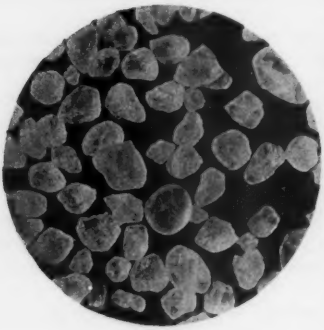

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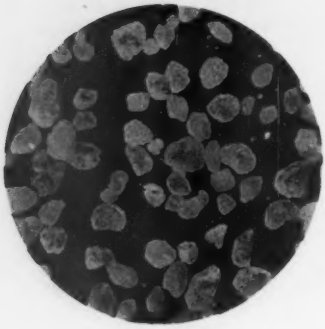

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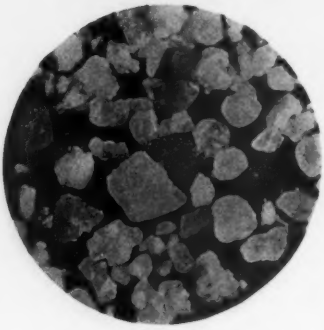


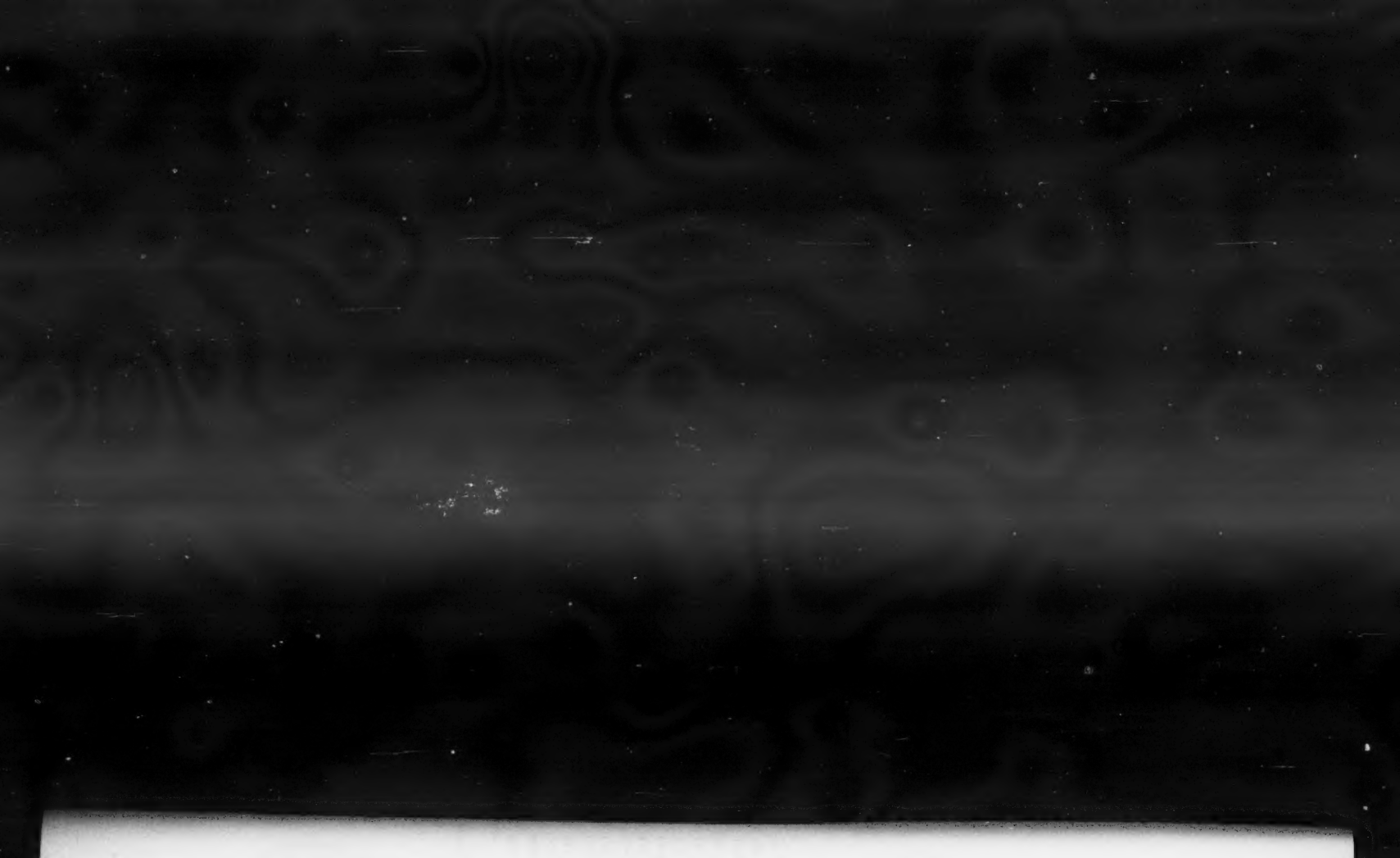
FIGURE 6

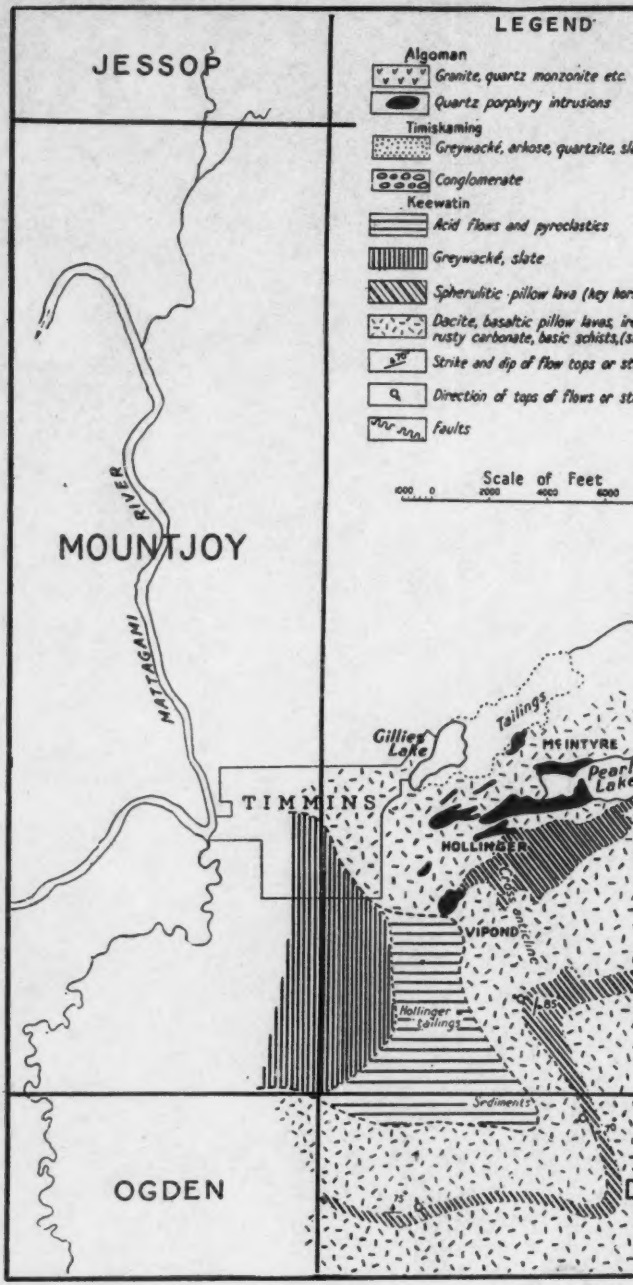
PHOTOMICROGRAPHS OF STAINED SURFACES AND SAND GRAINS

PLATE 4.—PHOTOMICROGRAPHS OF STAINED SURFACES AND SAND GRAINS

Figure

1. Stained surface of limestone. Black area—chalcedony of type shown in Pl. 3, fig. 1; pitted areas—green stained calcite; smooth light-gray area—unstained clay. Sample No. LD-40. ×20.
2. Stained surface of calcareous sandstone. Large light-colored area is a calcareous clay inclusion stained dark green. Sample No. LD-46. ×20.
3. Quartz grains from uppermost sandstone of the Morrison formation. Relatively uniform in size and moderately rounded. Sample No. LD-116. ×20.
4. Quartz grains from lower part of uppermost sandstone of the Morrison formation. Relatively uniform in size, subround to round, and coarser than upper part of same bed. Sample No. LD-115. ×20.
5. Quartz grains from lower part of second sandstone from top of Morrison formation. Variable in size and angular to subround. Finer-grained than uppermost sandstone of Morrison formation. Sample No. LD-114. ×20.
6. Quartz grains from sandstone bed containing dinosaur remains. Variable in size and angular to round. Speckled and dark-gray grains are chert. Sample No. LD-47. ×20.





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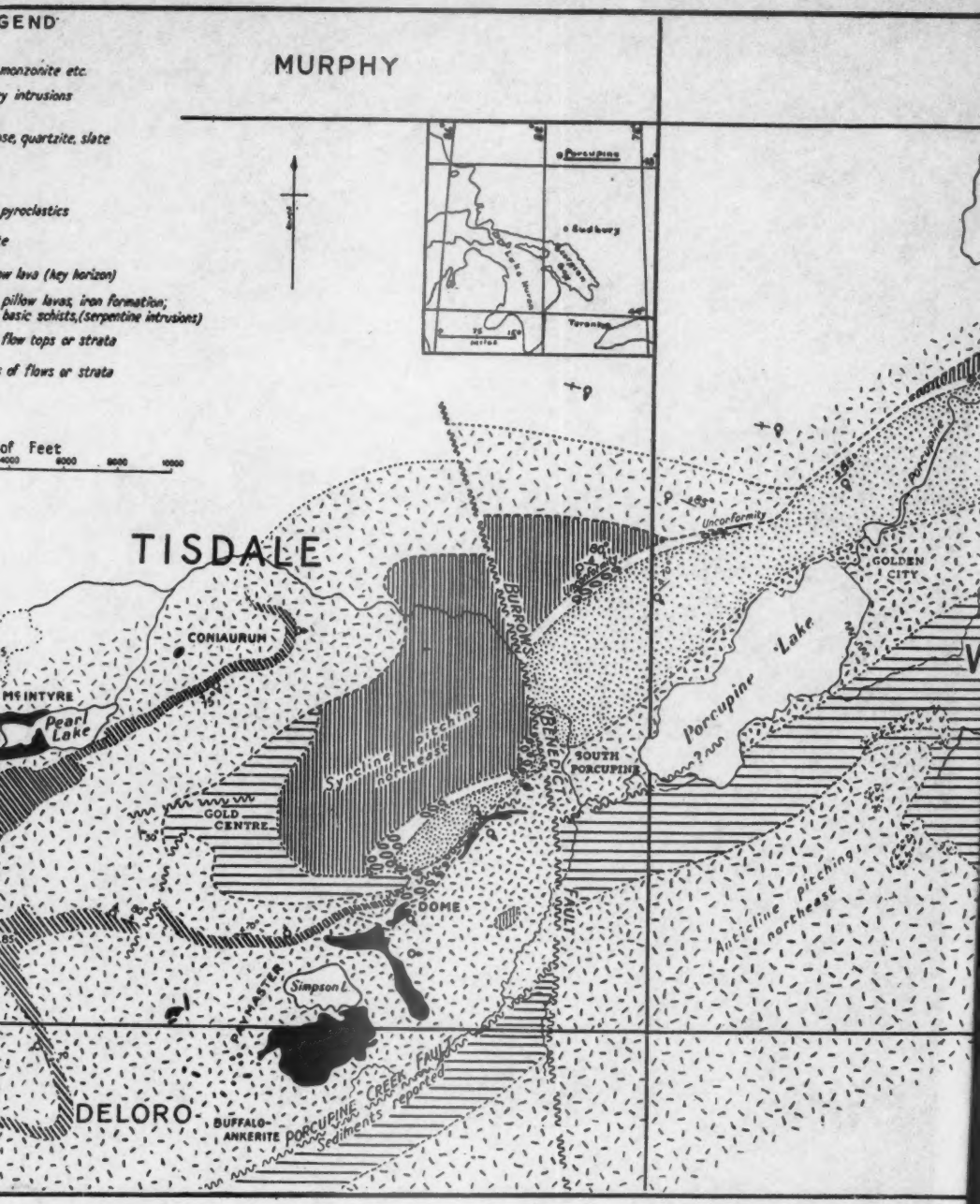

GENERALIZED GEOLOGIC MAP OF CENTRAL PART OF PORCUPINE DISTRICT

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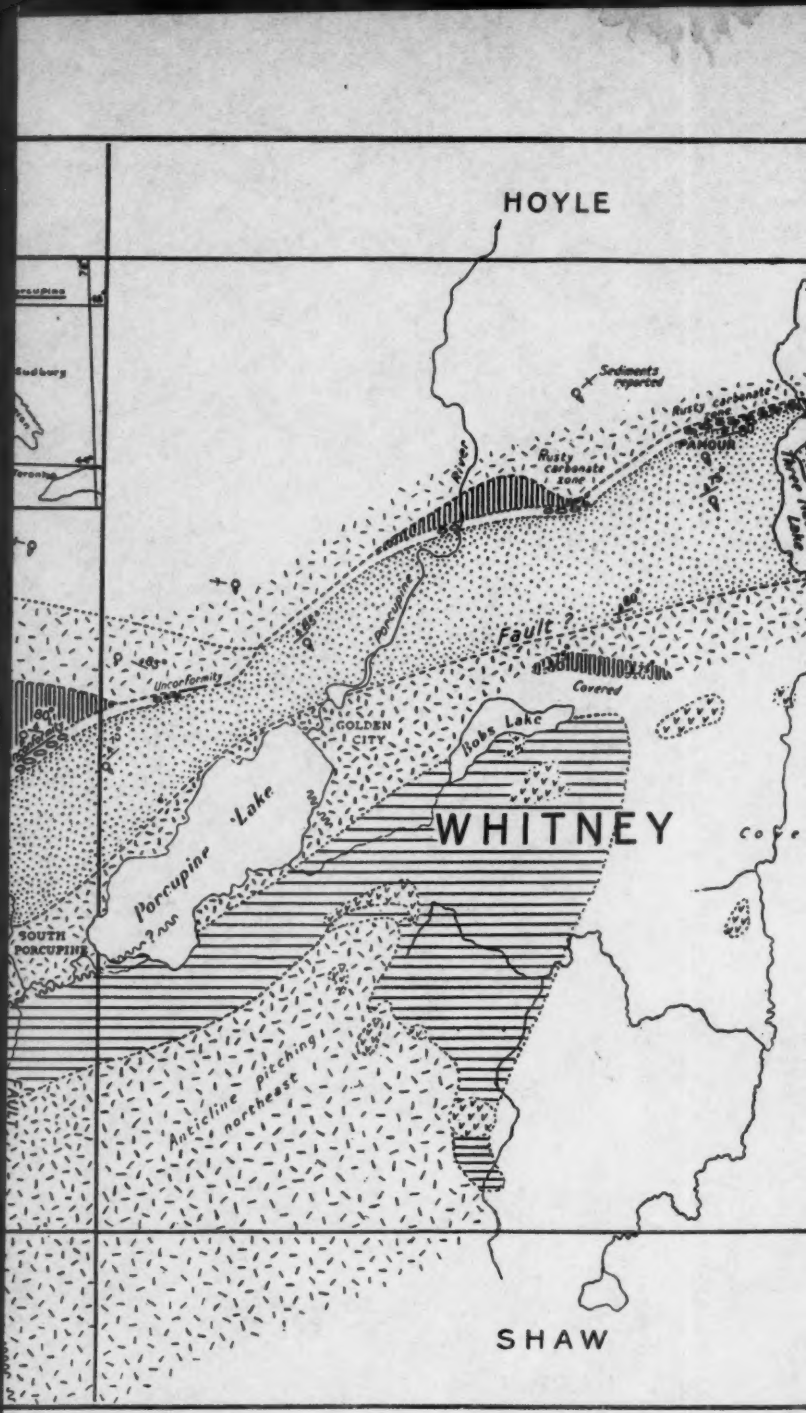

L PART OF PORCUPINE DISTRICT

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PORPHYRY OF THE PORCUPINE DISTRICT, ONTARIO

BY JAMES ERIC LLOYD EVANS

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ABSTRACT

The porphyry of the Porcupine district, Ontario, has been considered pre-Timiskaming by some and post-Timiskaming by others. A detailed study reveals that it is the result of two processes—igneous intrusion and metasomatic replacement—separated by a period of deformation.

Igneous intrusion is indicated by the stock- and dikelike forms crosscutting enclosing rocks and by their content of inclusions. Relative age is shown by the structures; porphyry crosscuts Timiskaming sedimentary beds but in most places contains a linear schistosity. Therefore, the intrusion followed deposition of the Timiskaming rocks but preceded the conclusion of the Timiskaming deformation. A Keewatin-Timiskaming unconformity indicates that the Keewatin rocks were folded and then eroded to a depth of at least 8000 feet before Timiskaming deposition.

Metasomatic replacement of the schistose intrusive is suggested by the lack of orientation of the feldspar and quartz "phenocrysts," their transection of the schistosity, and almost complete lack of strain. Similar features are common to "phenocrysts" in the porphyry of "ecole" structures, the porphyry fragments in Keewatin agglomerate, and the porphyry pebbles in Timiskaming conglomerate. In all these "fragmental" rocks, feldspar and quartz crystals, identical with those in the porphyry, occur in the matrix. Additional evidence of replacement is the presence in the stocks of lenses of carbonaceous porphyry which usually lie on the strike of carbonaceous schist lenses in the surrounding rocks. It is suggested that after the Timiskaming folding quartz and feldspar "phenocrysts" were deposited from permeating solutions. The Algoman granite is the probable source of the solutions.

INTRODUCTION

Among the intrusives of the Porcupine district the porphyry has excited the greatest interest because of its areal association with the gold-bearing veins. The views of geologists have differed greatly regarding the possibility of a genetic relationship between the porphyry and the gold-bearing solutions. Interpretations depend largely on the age of the intrusions relative to the surrounding rocks. The mineralization is generally thought to be post-Timiskaming (Burrows, 1924, p. 41–45; Graton *et al.*, 1933, p. 17; Hurst, 1936, p. 455). If the porphyry is likewise post-Timiskaming, as suggested by Burrows (1924, p. 30) and Langford (1941, p. 154), then a genetic relationship between it and the mineralization is possible. If, on the other hand, the porphyry is pre-Timiskaming, as contented by Graton and others (1933, p. 7), then the possibility of relationship is remote.

The porphyry occurs most commonly in stocks which are elongated in a horizontal plane along a southwest-northeast axis, following the directional trend of the regional schistosity, and plunge 40° to 60° NE. Less common are dikes up to 15 feet in width. Moreover, the porphyry is generally schistose and contains small-scale structural features identical with those of the surrounding rocks. Burrows (1924, p. 33) suggests that the schistosity was developed during an epoch of folding which occurred after the more intense Timiskaming deformation. Graton and others (1933, p. 6–7) postulate that the porphyry stocks were intruded prior to the deposition of the Timiskaming sediments and hence were involved in the folding of the Timiskaming. The facts so far stated are consistent with either hypothesis. The entire set of secondary structures could be assigned to one post-Timiskaming deformation, or it could be contended that the porphyry was intruded into the folded Timiskaming and deformed subsequently. Additional evidence will be presented to support the hypothesis that the porphyry was deformed with the Timiskaming rocks.

The mode of emplacement of the porphyry stocks has also been the subject of considerable speculation. Evidence for forceful injection appears to be lacking.

Stoping of the overlying rocks may have accompanied intrusion, but inclusions of wall rock are absent in most places; those found are small and confined to a narrow contact zone. Whitman (1927, p. 404–420) suggested a syntectic origin for the porphyry bodies. A number of features, such as the distribution of carbon and the porphyroblastic development of feldspar and quartz, support this idea.

The texture and color of the porphyry vary from place to place. Some variations are primary; others are due to later metamorphism. The primary variations may be related to mode of emplacement. For instance, one type of porphyry is dark, due to included carbon obviously derived from a carbonaceous schist very common in the region; the incorporation of carbon may have been due either to assimilation of schist or to syntectic development of porphyry from the schist. Some of the metamorphism of the porphyry can be attributed to alteration by the solutions which deposited the quartz veins, but most of it is caused by dynamic metamorphism modified by hydrothermal alteration.

A thorough genetic study of the porphyry of the Porcupine district must include, therefore, the determination of its relative age and mode of emplacement and an investigation of its metamorphism. The studies here reported were concerned with these problems. Five months in 1941 and 1942 were spent in the field covering an area of about 50 square miles, both on the surface and, where possible, underground. The porphyry was examined in detail, and particular attention was paid to its relationship with the surrounding rocks. In the field, structures of the rocks were studied insofar as they have a bearing on the age and origin of the porphyry. In the laboratory chemical, mineralogic, and petrographic methods were applied to the problems mentioned. The laboratory investigations were carried out at Columbia University during the winter of 1942–1943.

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GENERAL GEOLOGY

FORMATIONS

The general geology of the district has been adequately described elsewhere, so that only those features pertinent to the present investigation will be dealt with in detail.

The following table of formations is the one most generally accepted (modified, after Burrows, 1924; Hurst, 1936, p. 448):

Pre-Cambrian	
Keweenawan	Olivine diabase dikes.
Matachewan (?)	Quartz diabase dikes.
Algoman	Quartz monzonite and allied granitic intrusives.
	Albitite dikes.
	Quartz-feldspar porphyry dikes and stocks.
Haileyburian (?)	Serpentine.
Timiskaming	Conglomerate, graywacke, slate, arkose.
Keewatin	Graywacke, slate, tuff.
	Rhyolite, agglomerate, tuff, iron formation.
	Dacite, andesite, basalt.

KEEWATIN

The Keewatin is represented by a series of basic to acid volcanic rocks and a series of overlying sediments. The volcanic rocks are dacite, andesite, basalt, rhyolite, tuff, and agglomerate. They are altered to schist in many places. Many of the flows have retained their primary features, such as pillow structures, spherulites, and fragmental and ropy tops. These lesser features make certain horizons easily recognizable in the field and are keys to the major structures.

Stratigraphically above the series of flows is a thick agglomerate horizon (Fig. 1, Gold Centre and Hollinger tailings). This rock is peculiar in that from 50 to 90 per cent of the fragments mineralogically and texturally resemble the porphyry of the region. The remaining fragments consist of volcanic material, light-green schist, carbonaceous schist, and chert. In places the agglomerate resembles massive porphyry; only occasional foreign fragments and faint outlines indicate pyroclastic origin. For many years this rock was believed to be intrusive. Its exact stratigraphic position is not definitely known. The general assumption is that it lies immediately under the Keewatin sediments which occupy the center of the syncline west of the Burrows-Benedict fault (Pl. 1). Hurst (1936, p. 450) suggests that it may overlie the Keewatin sediments, overlapping them laterally.

The rocks south of the Porcupine Creek fault differ markedly from those to the north, and the exact relationship between the two sides is unknown. South of the fault the rocks consist of a thick series of tuff, agglomerate, acid flow rock, and intercalated bands of iron formation. Their age is uncertain, but it is generally

thought to be Keewatin as there are no known Timiskaming volcanic rocks in the region.

The youngest Keewatin rocks in the region are interbedded slate, arkose, and graywacke. These rocks have been highly metamorphosed, and fracture and flow cleavages have been developed in many places. Nevertheless, bedding, well shown by lithologic banding, has not been destroyed.

The Keewatin rocks contain abundant evidence of intense deformation. A planar schistosity occurs in them in some places; in others it is indistinct or lacking. A pronounced linear schistosity occurs in almost all the volcanic rocks north of the Porcupine Creek fault. This structure is well illustrated in the field by the tendency of the rocks to break into long, irregular splinters. Underground the backs of the crosscuts show slight schistosity, whereas the walls are extremely schistose. This is inferred to be due to the fact that the "pencils" of the schistosity are cut approximately at right angles to their long axes in the backs but parallel to them in the walls. Pillows in interbedded lavas have been deformed to crude cylindrical shapes (Pl. 2, figs. 1, 2). Fragments in flow tops show irregular outlines in a plane normal to the schistosity, whereas parallel to the schistosity they appear as pencil-like streaks. This feature is well illustrated in a glaciated outcrop on the south side of the "back" road to South Porcupine, about 200 feet east of the intersection with the road leading to the back gate of the Hollinger mine. The linear structure is also apparent in thin section. Many sections cut normal to the schistosity show no sign of foliation, but sections from the same specimens cut parallel to the schistosity show a decided parallel orientation in the long axes of the minerals and an aggregate extinction under crossed nicols.

TIMISKAMING

Conglomerate, slate, and arkose make up a group of sediments lying unconformably on the Keewatin rocks. Hurst (1936, p. 451) has tentatively correlated this group with the Timiskaming. These rocks have been intensely folded and contain most of the deformational features of the older rocks.

The base of the group is marked by a conglomerate about 100 feet thick. At various localities it lies in contact with different horizons in the Keewatin volcanics and sediments. A good exposure of such a contact is located about a mile north of South Porcupine. A few lenses of conglomerate also occur in the slate and graywacke. The conglomerate is poorly sorted. It contains pebbles and boulders of volcanic rocks, porphyry, granite, and diorite, and an occasional pebble of fine sedimentary material and banded iron formation. The pebbles and boulders range from less than an inch in diameter to several feet. At the Dome mine isolated pillows of lava occur as boulders in the conglomerate. Locally, the conglomerate is absent, and graywacke and slate lie against the volcanics, indicating (Hurst, 1936, p. 452), that these places were above the level of deposition when the conglomerate was laid down.

Above the conglomerate are interbedded slate and graywacke whose bedding is

emphasized by the contrast between the dark-gray slate and the lighter graywacke. Arkose and quartzite are associated with these beds in the eastern part of the region.

The deformation structures in the Timiskaming rocks are similar to those in the Keewatin sediments, so that these two groups cannot be separated on the basis of metamorphism. Both groups contain drag folds, flow cleavage, and fracture cleavage. Many readings were taken on the cleavages of the two groups; the cleavage in the Keewatin sediments usually strikes south of east, whereas the cleavage in the Timiskaming rocks strikes north of east. This suggested that cleavage strike might distinguish the rocks of the two ages. Enough exceptions were found, however, to render this method useless. Nevertheless, the observation does indicate that the rocks were deformed by separate orogenic movements and that the directions of compression of these two orogenies differed. Moreover, the similarity in degree of crystallization in rocks lithologically so alike suggests that the two orogenies were about equal in intensity, each producing recrystallization to a certain degree but unable to alter the rocks further. The pebbles in the conglomerate have been deformed in a manner similar to the elongation of pillows and fragments in the Keewatin rocks. Where the elongation is pronounced minute tension fractures cut the pebbles normal to their long direction.

HAILEYBURIAN

Serpentine occurs in large masses in various parts of the area and is presumably the result of hydrothermal alteration of basic intrusives, though some of it may have been derived from basic flow rocks. Veinlets of asbestos and talc cut the serpentine in a few places. This rock is clearly older than the Algoman granite, which is intrusive into it, and hence is tentatively correlated with the Haileyburian group of intrusives.

ALGOMAN

Quartz feldspar porphyry.—The porphyry masses are confined chiefly to four localities. In the vicinity of Pearl Lake the porphyry occurs in a group of large stocks and a few small dikes. The largest stock is the Pearl Lake mass, which is approximately 5000 feet long and 1500 feet wide. More is known about the occurrence of porphyry in this locality than elsewhere in the camp because of the exposures in the Hollinger, McIntyre, and Coniaurum mines. Around Simpson Lake the porphyry is partially exposed in the underground workings of the Dome, Preston East Dome, and Paymaster mines. The largest single mass of porphyry lies in the northeastern corner of Shaw township, where it forms Mount Logano. A number of small dikes both south and west of Mount Logano may be related to this mass. The remaining exposures of the porphyry are in the east-central part of Ogden township, in small stocks and dikes. Several have been intersected in the Nabob and De Santis mines.

The form of the porphyry stocks reflects the general regional structure. Most of them are ellipsoidal in plan with their long axes pointing northeast. The only notable exception is west of the Preston East Dome mine, where a mass points northwest. The sides of the porphyry masses are fairly regular and are parallel

FIGURE 1. PILLOW LAVA ON NEARLY HORIZONTAL OUTCROP

FIGURE 2. ELONGATION OF PILLOW STRUCTURE ON VERTICAL OUTCROP
Hammer rests on horizontal section across a "tubular" pillow. Figure 1 shows top view of "knob" at left.

FIGURE 3. INTERSECTING FRACTURES IN PORPHYRY, SEALED BY TOURMALINE
View shows roof of crosscut, and pipe trends approximately N-S.

OUTCROPS

FIGURE 1. PORPHYRY (TOP) IN CONTACT WITH TIMISKAMING SEDIMENTARY ROCKS (BELOW)
Bedding of sediments crumpled at contact and for a short distance away from it.

FIGURE 2. BANDED PORPHYRY IN HOLLINGER MINE
Coin (center) is large Canadian five-cent piece. (Photograph by C. H. Wilkins).

FIGURE 3. ECOLE STRUCTURES
Elongated parallel to schistosity. Note schist fragments in matrix. Arrow indicates a fragment. (Photograph by C. H. Wilkins).

PORPHYRY UNDERGROUND

to the strike of the enclosing rocks and to the regional schistosity. The ends of the porphyry stocks, on the other hand, are extremely irregular and dovetail into the wall rock. Some of these projections from the stocks are dikelike and extend into the country rock for several hundred feet. In addition the porphyry stocks pitch 40°–60° toward the east or northwest; they appear to widen slightly at depth.

The porphyry contains a high proportion of sericite and iron carbonate in some places; weathering of the carbonate produces a rusty coloring. It has a pronounced planar schistosity, particularly around Pearl Lake. In fresh rock, the planar schistosity is not so prominent, but the rock has the linear schistosity and splintery fracture mentioned in connection with the Keewatin rocks. This linear structure is well illustrated by inclusions of wall rock in porphyry. One such inclusion was at least 3 feet long and not more than 3 inches across. The long axis of the inclusions corresponds with the direction of linear schistosity. In addition, they contain linear schistosity parallel to that in the surrounding material. Alignment and elongation are also apparent in thin sections. Sections cut parallel to the lineation show a parallel orientation of sericite blades; those cut normal to the lineation show no apparent foliation.

The fresh porphyry is a massive rock containing abundant quartz and feldspar "phenocrysts" up to a quarter of an inch across. Color ranges from dark gray and black to yellow or pink. In general the lighter color accompanies greater schistosity and appears to represent greater alteration. All porphyry observed shows a linear structure of varying intensity. Thin sections reveal "phenocrysts" of oligoclase ($Ab_{85}An_{15}$) and of quartz in a finely granular groundmass of small, interlocking grains of quartz and feldspar. Sericite is present in all specimens but in varying amounts. It shows no orientation in some sections cut normal to the schistosity; in others it shows a slight preference for two intersecting directions making an angle of approximately 45° and thus similar to the shear directions in stressed rocks. In sections cut parallel to the schistosity sericite occurs in parallel blades with aggregate extinction under crossed nicols. Leucoxene is a common accessory in the porphyry. A concentrate of this material, which is opaque to transmitted light and white or light yellow in reflected light, gave an X-ray pattern identical with rutile. The leucoxene occurs as scattered spots and streaks and preserves outlines of pre-existing minerals (Pl. 4, fig. 1). Hornblende and sphene were found in one specimen. Chlorite is present in some of the porphyry but is not common. Pyrite occurs in most porphyry specimens but is more abundant in altered porphyry adjacent to veins. Tourmaline occurs commonly in the lighter-colored, more schistose porphyry and cements some of the fractured porphyry.

Granite.—Granite in the form of small stocks, tongues, and dikes occurs in the southern and eastern parts of the area. The largest exposures are east of Porcupine Lake. The granite is light pink and even-textured, with minor biotite and hornblende. In contrast to the schistose porphyry, it is unfoliated and unsheared in most places. In a few places breccia cracks are filled with fine-grained chloritic material. Quartz veins are numerous. Large tourmaline-quartz veins, similar to those found in the mines, cut the granite on the Two-in-One property in south-central Whitney township.

MATACHEWAN AND KEWEENAWAN

Many diabase dikes ranging in thickness from a few inches to 150 feet intrude all other rocks. Two types are distinguished, olivine and quartz diabase. The quartz diabase dikes strike toward the northwest and are generally believed to be Matachewan in age; the olivine diabase dikes usually strike more nearly due east and are probably Keweenawan.

STRUCTURAL HISTORY OF THE PORCUPINE DISTRICT

The Keewatin and Timiskaming rocks of the Porcupine district have been tightly folded. The folds trend northeast-southwest and plunge toward the northeast. An apparent convergence of the fold axes immediately southeast of the town of Timmins has been explained as a cross anticline by Graton and others (1933, p. 8) and Hurst (1936, p. 454).

At least two periods of folding are recorded. The first involved the Keewatin rocks before the deposition of the Timiskaming sediments; the second folded the Timiskaming rocks but preceded the intrusion of the Algoman granite. For purposes of description the first period will be referred to as the Keewatin folding, and the second as the Timiskaming. Opinions vary regarding the intensity of the Keewatin folding, but most assume a rather mild deformation. Structural study was made to determine the intensity of this folding.

Timiskaming rocks unconformably overlie Keewatin sediments in an outcrop a mile due north of South Porcupine. The Keewatin sediments strike N. 80° W. and dip 68° N., and the Timiskaming rocks strike N. 60° E. and dip 75° SE. A well-developed fracture cleavage in the Keewatin sediments strikes N. 40° W. and dips 42° NE., indicating overturning to the south. Flexing of these beds in folding affects the limited area of a contact, such as between the Timiskaming and Keewatin rocks, much like simple tilting. An approximation as to what the contact looked like before flexing may be obtained by graphically tilting the Timiskaming rocks back to a horizontal position. This was done by means of stereographic projection, which gave a pre-Timiskaming strike for the Keewatin beds of N. 73° W. and a dip of 54° S. The strike of the Keewatin fracture cleavage, similarly calculated, was N. 78° W., and the dip 85° S. These two computed attitudes indicate that normal folding, with the axial plane essentially vertical, preceded deposition of the overlying Timiskaming. In applying this method, certain qualifications must be recognized. For example, such deformation involves a three-dimensional change in shape. In this area, the folding resulted in a maximum elongation in the intermediate direction of stress (the *b* axis of petrofabrics). Normally elongation would cause an apparent oversteepening of the contact. In this case, however, the direction of elongation (upward 40–60° in a S. 60° W. direction) comes within 10°–20° of lying in the planes of both the bedding and fracture cleavage in the Keewatin rock, so that the effect of elongation must have been minimized. The change in attitude inferred from the assumption of tilting may, therefore, be correct in principle though not determinable quantitatively.

From the same outcrop the amount of erosion before the deposition of the Timiskaming rocks was estimated. The basal conglomerate contains pebbles of volcanic

rock and granite, indicating that erosion had progressed into the underlying intrusives. A line drawn from this outcrop to a point 9000 feet to the north passes over a succession of Keewatin sedimentary and volcanic rocks in which no beds are repeated. All the horizons in this succession abut against the Timiskaming rocks to the east. Allowing an average dip of 70°, which corresponds roughly to that now observed, a calculated thickness of 8000 feet of Keewatin rocks was eroded before the deposition of the Timiskaming conglomerate.

Additional evidence regarding the Keewatin deformation lies in the minor structures in the rocks. Divergent strikes of the cleavages in the Timiskaming and Keewatin sedimentary rocks have been mentioned. No uniformity in strike of schistosity was found in the Keewatin volcanic rocks, but in general it is east. In a few places the volcanic rocks contain two schistosities. The best example is on the side of the road about 1½ miles south of the South Porcupine station. Here one schistosity striking N. 25° E. and dipping 85° SW. is cut by a second striking N. 55° W. and dipping 35° SE. Both are well developed and indicate two deformations.

Thus the Keewatin rocks were tilted to a calculated maximum angle of 54°, and at least 8000 feet of rock was eroded before the deposition of the Timiskaming sediments. Folding produced fracture cleavage in the Keewatin sediments and probably flow cleavage in the Keewatin volcanic rocks.

Although the Keewatin deformation was probably fairly intense, most of the deformation structures now seen in the rocks of the district can be assigned to Timiskaming folding. The peculiar deformation of pillow lava, fragments in flow tops, and inclusions in porphyry, as well as the widespread linear schistosity, can all be related to the Timiskaming folding. The most direct evidence of this is that the pebbles of the conglomerate in the Dome mine are also drawn out to pencil-like shapes. In addition, the porphyry, which is post-Keewatin, shows these structures and therefore suggests that they were formed during the Timiskaming disturbance.

No deformation of comparable magnitude has disturbed the rocks since the Timiskaming folding. This is indicated by the massive nature of the granite of the region. The granite was fractured prior to the introduction of the quartz veins, but under the stress which produced the fractures the rocks acted as brittle media, whereas under the more intense Timiskaming stress the rocks flowed plastically. About 100 miles south of the Porcupine district the Timiskaming rocks are overlain unconformably by a thick series of Cobalt sedimentary rocks. These rocks are moderately tipped to dips not exceeding 30°, supporting the idea that no great disturbance occurred in this region after the close of Timiskaming time.

The above evidence permits an interpretation of the relative ages of the porphyry and the quartz veins. Deformational features present in the porphyry but absent in the granite show that the porphyry is older than the granite. Quartz-tourmaline-ankerite veins cut the porphyry in all the mines of the district. Similar veins also cut the granite, indicating that this type of mineralization is later than the granite.

AGE OF THE PORPHYRY

There has been some question as to whether the porphyry intrusions are pre-Timiskaming or post-Timiskaming. The relationship of the porphyry to the

Keewatin and Timiskaming deformations furnishes a clue to its relative age. The conformity of the porphyry stocks to the general regional structures indicates that the porphyry was intruded into folded Keewatin rocks. It has been shown that the linear schistosity in the porphyry was formed by Timiskaming deformation. Therefore, the porphyry intrusions were injected after the Keewatin deformation and before the deformation that followed Timiskaming sedimentation.

Porphyry lies in contact with the Timiskaming sedimentary rocks at the Dome mine. At a contact on the 17th level the porphyry cuts across several distinct beds including a conglomerate horizon. The contact is quite sharp and shows little evidence of disturbance other than a slight wrinkling of the sedimentary beds near the contact (Pl. 3, fig. 1). The contact itself is very irregular, and small round knobs of porphyry project into the sedimentary beds. A suggestion has been made that such contacts between porphyry and Timiskaming sediments are fault contacts. In the opinion of the writer, the contact is too tight and to irregular to admit of faulting on the scale required to bring the porphyry into its present position against the sediments.

If the above evidence is accepted, the porphyry intrusion can be dated more exactly. It must have been injected after the deposition of the Timiskaming sedimentary rocks and before the cessation of Timiskaming deformation. The intrusion could have taken place during the latter part of the folding, in which case the porphyry may be genetically related to the Algoman granite—that is, the porphyry may be the first expression of widespread igneous activity and may have been followed, after deformation, by the main granite. Such a complex history of intrusion coupled with great orogeny is not as surprising as was formerly believed. It has been described from many regions (Lovering and Goddard, 1938, p. 30–34; Peach and Horn, 1930, p. 192–197; Freeman, 1934, p. 44; Hinds, 1934, p. 182–192), especially where the structure is less complex and the rocks less altered so that the record is clearer.

ORIGIN OF THE PORPHYRY

ALTERNATIVE HYPOTHESES

The absence of any evidence of forceful intrusion is a peculiar feature of the porphyry stocks. It is possible, though unlikely, that much of the evidence was destroyed during the deformation that followed intrusion. The porphyry stocks cut across all discernible horizons in the rocks in which they are found. There is no evidence that the porphyry made room for itself by pushing the rocks aside, nor does the schistosity flow around the porphyry. Since the schistosity passes from the wall rock to the porphyry, it may have developed after intrusion. The bedding in the Timiskaming sedimentary rocks is slightly crumpled at its contact with the porphyry, but this is ascribed to the folding which involved both rocks. The interfingering of the porphyry and the wall rock off the ends of the stocks suggests injection, but no displacement of horizons in the wall rock was detected.

Progressive solution of the wall rock by the magma may have been the mode of intrusion. In many places along the contacts the porphyry contains inclusions of

wall rock; in others it is contaminated by carbon derived from carbonaceous schist; in still others the contact between wall rock and porphyry is gradational. The interfingering of porphyry with wall rocks suggests a quiet advance of a fluid magma rather than solution.

The lack of evidence of forceful intrusion suggests a metasomatic origin. Grout (1941, p. 1528–1530) points out the difficulties of proving a metasomatic origin for rocks which look igneous. In the present case it is not possible to develop and apply all the criteria suggested by Grout. Enough features are present, however, to justify inferences as to the origin of the porphyry.

EVIDENCE FOR IGNEOUS INTRUSION

Some features suggest that the porphyry was an igneous rock which was intruded into the folded Keewatin. In most places it looks like an igneous rock, and, although close inspection reveals that the rock is schistose, the feldspar and quartz crystals in it bear no relationship to the schistosity. The stocklike and dikelike shapes of the porphyry masses also resemble igneous intrusions. Further, the porphyry masses crosscut bedded horizons, and in a number of places the contacts are quite sharp.

Angular inclusions of country rock in the porphyry along its margin strongly suggest an intrusive origin. The inclusions resemble adjacent, unbrecciated flow rocks, and the schistosity in them is parallel to that in the porphyry. Many of them are angular in a plane normal to the schistosity but linear parallel to it indicating that they existed as fragments and were elongated during the deformation. Therefore, they were probably enclosed in a pre-deformation intrusive and are not unreplaced remnants of the later metasomatism which formed the quartz and feldspar metacrysts. This suggests that intrusive bodies may have been present in the porphyry localities before the Timiskaming deformation.

EVIDENCE FOR METASOMATIC ORIGIN

General statement.—The present case, though believed by the writer to be essentially one of metasomatism following intrusion, differs from most of those in the literature in that the source rock of the metasomatizing solutions is nowhere visibly in contact with the porphyry; the source rock, therefore, can only be inferred.

Quartz and feldspar phenocrysts.—The feldspar "phenocrysts" of the porphyry range from small individuals with amoeboid shapes to large, well-formed crystals. The feldspars are fresh and well preserved in many specimens. They contain abundant inclusions of quartz, sericite, carbonate, and leucoxene. A few bear inclusions of amorphous carbon. Some of the "phenocrysts" have been fractured, and the fractures are filled with sericite or rehealed by feldspar of the same composition as the phenocrysts (Pl. 4, fig. 2). Many of the feldspars have clear rims, pointing to secondary growth (Pl. 4, fig. 3). The feldspars in the more schistose porphyry are fractured, and sections cut parallel to the schistosity reveal the drawing out of some of the "phenocrysts"; even in such rocks the feldspars have been regenerated, and the fragments more or less recemented by secondary feldspar (Pl. 4,

figs. 4, 5). In the same sections feldspars with amoeboid outlines show no relationship to the structures in the groundmass, thus suggesting secondary growth after deformation. Feldspar in most of the fresh prophyry seems to be unrelated to structures in the groundmass. The individual feldspar crystal is not oriented parallel to any recognizable direction of elongation in the rock; it transects the schistosity as though a patch had been cut out of the groundmass and a feldspar of identical size and shape inserted in its place (Pl. 4, fig. 6). Van Hise (1904, p. 702) observed similar relationships of porphyritic minerals developed in schist, and they have been reported in many places and ascribed to secondary origin. In a few places the sericite blades curve slightly around the feldspar as though it had pushed them aside. This feature is exaggerated in thin sections which contain streaks of carbon or leucoxene. The streaks follow the schistosity to the feldspar grain where they diverge around it. Tongues of fresh feldspar project into the groundmass from some of the crystals (Pl. 5, fig. 1). These tongues are short, however, and not nearly so well developed as the tongues from the quartz.

The quartz "phenocrysts" are even more distinctive than those of feldspar. The shapes are variable, but most crystals are rounded. Many of the crystals are deeply embayed, and others have enclosed groundmass material and feldspar (Pl. 5, fig. 2). Inclusions of indeterminable particles arranged in chains and clusters are common. Some of the grains have a margin of clear quartz rimmed on the inside by a row of inclusions suggesting either regrowth or incorporation of groundmass during growth (Pl. 5, fig. 3). A remarkable feature of the quartz is the almost complete lack of strain and fracturing. This feature, coupled with the fact that quartz cuts across the schistosity, and the other facts given above all indicate that the quartz formed later than the enclosing groundmass material.

One specimen of porphyry contains hornblende and sphene. The specimen is identical in appearance with other fresh porphyry which lacks these two minerals. It is dark gray and contains prominent white feldspar crystals ranging from a sixteenth to a quarter of an inch in diameter. A schistose structure is apparent in the hand specimen, but the feldspar crystals are not oriented. In thin section the resemblance to normal porphyry is very striking. The groundmass is fine-grained and schistose with the feldspars transecting the schistosity. Some of the feldspars have clear rims indicating secondary growth; others have small fractures rehealed by feldspar with the same orientation and composition as the host. Hornblende occurs as small crystals, a few of which have good crystal outlines resembling those preserved by leucoxene in normal porphyry. In contrast to the feldspar, the hornblende shows a tendency toward orientation, the long direction of the crystals roughly corresponding to the direction of the schistosity. Sphene occurs as small wedge-shaped crystals and as irregular patches in the groundmass. Both hornblende and sphene have been included in large feldspar crystals, indicating that the feldspar developed after these minerals.

Textural and mineralogical variations.—Textural variations are very striking within the porphyry bodies. The more massive, coarse-textured porphyry is fairly common in certain porphyry stocks and particularly those south of the Pearl Lake mass. All textural changes in the porphyry are gradational. Feldspar "pheno-

crysts" appear in clusters in certain places. One part of the northern porphyry, exposed on the 4175 level of the McIntyre mine about 1400 feet north of 4167 drift, contains large blue-gray feldspar crystals in a yellow schistose groundmass, giving an extremely patchy appearance to the porphyry. Some of the porphyry lacks feldspar. This rock is generally the most schistose and is identified as porphyry chiefly by its light color and the presence of quartz "eyes." Under the microscope the quartz "eyes" appear to be metacrysts. The shape of the grains varies from irregular to rounded, and none show strain. The edges of the grains project like tongues into the schistosity of the groundmass (Pl. 5, fig. 4). In many cases it is almost impossible to determine exactly where the quartz stops and the groundmass begins. Similar features have been described from undoubted secondary accretions of quartz in typical metamorphic rocks (Behre, 1933, p. 178; Harker, 1939, p. 242). Inclusions are abundant and support the idea of metasomatic origin for the quartz.

The texture of the groundmass in the porphyry also shows great and abrupt variations. In many sections it is so fine that the components are indeterminable; in others it is noticeably coarser; and in still others there is a mixture of finer and coarser groundmass.

Chlorite, though not common in the porphyry, occurs in a number of places and contributes characteristic textural features to the rock. It is particularly abundant in the porphyry of the ecole structures and in some of the porphyry of the Preston East Dome mine. The porphyry of the ecole structures will be described later. The chlorite-bearing porphyry of the Preston East Dome mine is a fine-grained, green rock with marked linear schistosity. It contains well-formed feldspar crystals which in thin section are observed to be composed of aggregates of small interlocking grains. The component parts of the crystal extinguish and reach maximum brightness together under crossed nicols, but at any intermediate point the composite nature of the crystal is apparent (Pl. 5, fig. 5). Goodspeed (1937, p. 1134–1135) has described similar features for feldspar porphyroblasts in hornfels from the Wallowa Mountains of northeastern Oregon. The groundmass of this porphyry contains large flakes of light-green chlorite in fresh, small, interlocking grains of quartz and feldspar. Chlorite occurs also as irregular flakes in the feldspar (Pl. 5, fig. 6).

Chlorite is present also in the finely banded porphyry of the Preston East Dome mine. The banding as seen on a rock face underground is moderately continuous and resembles sedimentary layering. It is due to the concentration of chlorite in layers about a thirty-second of an inch thick and about an eighth of an inch apart. This porphyry occurs as dikes in or adjacent to a highly foliated chloritic rock. One such dike has been traced directly into the main mass of porphyry in the underground workings of the mine. In thin section the groundmass of the porphyry appears coarser than that of the normal porphyry of the region and is composed of interlocking grains of quartz and feldspar, slightly elongated in the direction of foliation. Parallel blades of chlorite are very abundant. "Phenocrysts" of quartz and feldspar are relatively scarce, but those present are irregular in shape and bear no relationship to the groundmass schistosity. The adjacent wall rock resembles the porphyry, but the wall rock contains some biotite and chlorite in greater amounts,

and it lacks "phenocrysts". It is evident from the thin sections that the wall rock and the groundmass of the porphyry have undergone the same deformation but that the feldspar and quartz "phenocrysts" developed after the deformation.

Like the great variations of the porphyry in texture, the variations in color are not normal for an intrusive rock. In general, the more schistose the porphyry, the lighter is its color and the fewer are the feldspars contained. This would indicate that the feldspar crystals had been destroyed in the more schistose porphyry. On the other hand, the feldspar and quartz crystals in the fresher porphyry have been shown to have developed after the schistosity. The secondary growth of the feldspars and the healing of feldspar fractures by feldspar of the same composition show that solutions depositing feldspar were present in the rock or entered it after it was solid. It is concluded, therefore, that all the rock within the outlines of the present porphyry masses was schistose to much the same degree, but that the later sporadic development of quartz and feldspar crystals partially obliterated the schistosity. Where the crystals developed in great numbers, the schistose character of the rock was modified to a great extent; conversely, where the crystals were more widely separated, the rock retained its schistose character and altered appearance.

Carbonaceous porphyry.—Another striking variation in the porphyry is a dark-gray phase, due to varying amounts of included amorphous carbon. In some localities dark porphyry is common which, aside from color, is similar to the normal fresh porphyry in all respects. The quartz and feldspar crystals in most specimens of the dark porphyry are black. A few specimens, however, contain white feldspars, and others have white feldspars with black centers. The feldspar crystals in some of the black porphyry contain more carbon than is present in the groundmass. Carbon is smeared throughout the feldspar or is aggregated into small angular, unoriented clumps. Many feldspar crystals contain carbon in the center and have a clear rim; still others in the same section are entirely free from carbon. All specimens of dark porphyry show a slight linear schistosity.

Dr. C. P. Jenney of the McIntyre Porcupine Mines states that the carbonaceous porphyry usually occurs as lenses in the main porphyry, lying directly on strike from the carbonaceous schist. This statement has been confirmed by detailed observations made by the writer. Carbonaceous zones extend into the porphyry bodies for several hundred feet in many places. Examples may be seen on the 4175 level of the McIntyre mine and on the 3200 level of the Hollinger mine. The contacts of the black porphyry with normal porphyry are sharp in most places, and the gradation between the two rarely extends over a distance of more than 2 or 3 feet. The gradation in some places takes the form of banded porphyry.

Thin sections of black porphyry reveal that the carbon is scattered throughout but tends to segregate into streaks parallel to the schistosity in some sections and at least partly parallel to the foliation in others. A coarser banding of the porphyry was observed in both the McIntyre and Hollinger mines. Underground it shows as a series of light and dark band and lenses of varying widths (Pl. 3, fig. 2). This banding is due to carbon, which is more concentrated in the darker bands. The schistose structure in this rock cuts the banding at an angle (Fig. 1), clearly indicating that the schistosity was superimposed on the banding. In thin section the

black porphyry contains irregular streaks of carbon which fade out at the contact with the lighter porphyry. These streaks are aligned roughly with the schistosity. Neither megascopically nor microscopically is there a textural change between the light and dark bands. In neither the light nor the dark phase are the crystals oriented; in both they cut across the schistosity in the rock, apparently post-dating the development of the schistosity.

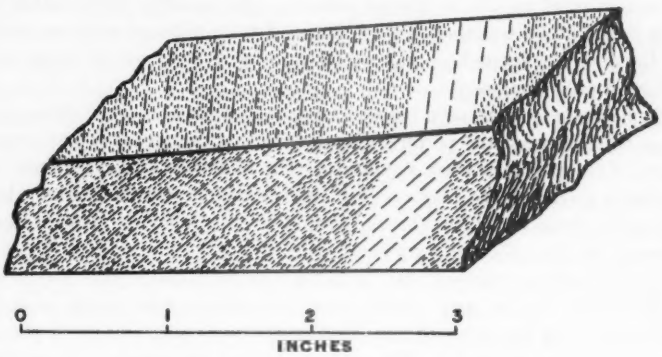

FIGURE 1.—*Sketch of a specimen of banded porphyry*
Showing schistosity cutting banding. Schistosity is indicated by dashed lines.

Graton *et al.* (1933, p. 5) suggest that the carbon of the carbonaceous schist was syngenetic. This is the best explanation presented so far, as no evidence of introduction of carbon was observed. In any case, the carbon was present in the schist before the formation of the porphyry, as evidenced by carbon-rich inclusions of this schist in it. It is probable therefore, that the carbon of the carbonaceous porphyry was derived from the schist. The banded porphyry clearly shows that the carbon antedates the development of the schistosity. If the carbon were introduced after the development of the schistosity, the schistosity and banding should be parallel. The quartz and feldspar "phenocrysts" appear to have formed after the schistosity. Thus the evidence points toward the metasomatic development of quartz and feldspar crystals in a schistose rock containing carbon. Additional evidence lies in the fact that carbonaceous porphyry occurs not only in contact with carbonaceous schist but also as lenses in the normal porphyry on the strike of carbonaceous schist. On the other hand, the presence of angular fragments of carbonaceous schist in the porphyry in other places indicates incorporation of schist in fluid magma.

Ecole structures.—Whitman (1927, p. 405–415) described curious phenomena along the borders of some of the porphyry bodies, and almost identical phenomena have been described in the lower Keechelus of the Cascade Range of Washington by Goodspeed and Coombs (1937, p. 12–23). Whitman observed "elongated lenticules of porphyry" in the greenstone adjacent to its contact with the stocks. He called these phenomena "ecoles" because of their collective resemblance to schools of fish. The nature of the ecole structure is not too apparent on the outcrops now exposed in the district. Many of the outcrops described and photographed by

Whitman have since become overgrown or covered by debris from the abandoned Porcupine Crown mine. Ecole structures were observed in a number of small outcrops on the southwest corner of the Hollinger property. The dimensions of the lenses range from less than an inch to about 4 by 9 by 36 inches. The majority of them at this place, however, are small and vary from elliptical to angular when seen in the nearly horizontal plane of the surface outcrops. The porphyry lenses stand out in relief on the weathered surface. The material surrounding them is chlorite schist, and the contact is very sharp. On the same outcrop are small areas where the porphyry lenses become so numerous that they appear to merge and form a matrix for schist fragments.

Ecole structures can be seen more clearly underground, as the surfaces are fresh and the relief between the porphyry lenses and their matrix does not obscure the contact. An exposure of ecole structure on the 3200 level of the Hollinger mine is particularly good because here one can see complete gradation within about 100 feet from the ecole structures through porphyry containing abundant inclusions to massive porphyry. The porphyry lenses consist of dark-gray, coarse, fresh porphyry in which the feldspar "phenocrysts" attain a maximum diameter of a quarter of an inch (Pl. 3, fig. 3). A slight linear schistosity is apparent in the rock, but the "phenocrysts" are not oriented nor do they show any relationship to the groundmass structure (Pl. 6, fig. 1). The contact between the porphyry lenses and their matrix is usually sharp, but in a few places it is gradational over a distance of about an inch. Feldspar "phenocrysts" not only occur in the lenses but also cut across the contact between the lenses and their matrix; moreover, such "phenocrysts" are found as isolated crystals or groups of crystals in the matrix. The matrix of the porphyry lenses is in some places fine chlorite schist; in others it is composed of obviously fragmental material (Pl. 7, figs. 3, 4, 5). Fragments range in size from an eighth of an inch to an inch and are composed of a variety of materials. In an area of about 9 square feet the writer counted 8 types of intermingling fragments, all of which resemble typical volcanic rocks of the region. In addition, fragments of carbonaceous schist are very abundant. All fragments are drawn out into pencil-like shapes conforming to the regional linear structures.

Nearer the main porphyry stocks the porphyry lenses are more numerous but smaller; still nearer, the lenses disappear, and the rock consists of abundant, closely packed fragments in a porphyry matrix. The latter rock is in sharp contact with the porphyry of the stock, which lacks inclusions. The porphyry of the lenses, that which forms the matrix for the inclusions, and that of the porphyry stocks are so similar in texture and composition that there can be little doubt of their common origin. The composition of the feldspar crystals in the ecole porphyry, the ecole matrix, and the porphyry stocks was determined accurately by oil-immersion methods and were found to be identical.

Whitman (1916, p. 256) first ascribed the origin of the ecole structures to injection of a fluid magma in a series of small pipes. Later he (1927, p. 415–420) postulated the development of porphyry in place as a syntectic magma. Graton *et al.* (1933, p. 6) contended that the ecole structures represent a breccia formed during or after intrusion of the porphyry. The bulk of the evidence appears to favor the syntectic

origin. The origin of the ecole structures through brecciation seems impossible because: (1) The ecole structures have a variety of intermingling fragments associated with them. (2) These fragments have suffered intense deformation, whereas the porphyry of the lenses was not involved in any deformation, as shown by the absence of recognizable orientation and fracturing in the feldspar and quartz "phenocrysts". (3) Feldspar crystals were developed in the matrix surrounding the lenses after this material is believed to have been deformed.

The ecole structures appear to have been formed by two intersecting shears in some places (Pl. 3, fig. 3). W. H. Bucher (personal communication) suggests that the fragmental rock was originally bedded and that the shearing had broken the more competent beds into rhombic pieces which were further rounded by more shearing. A similar explanation was offered by C. H. Behre (personal communication) who suggests that bedding is not essential to produce the structure; he ascribes the lenticular form to rounded rhombs produced by intersecting shear planes. The writer saw no evidence of bedding in the adjoining unreplaced fragmental rock. In any case it is agreed that the shearing antedated the formation of the feldspar crystals. The explanation offered here for these structures is that a fragmental rock, presumably agglomerate, was selectively attacked by solutions which deposited quartz and feldspar. It is unlikely that the porphyry lenses were ever fluid because the groundmass of the porphyry is schistose, whereas the "phenocrysts" show no relationship to this structure.

The agglomerate.—The porphyry of the Keewatin agglomerate is so similar in appearance to that found in the porphyry stocks that it was felt necessary to include it in the present study. The distinctive features of the agglomerate previously mentioned are the preponderance of porphyry fragments and the massive, unfragmented appearance of the rock in many places.

Contacts between porphyry fragments cannot be discerned in thin sections as the porphyritic texture is continuous across such boundaries. In thin sections the porphyry of the fragments is almost indistinguishable from that of the stocks. It contains numerous well-formed crystals of feldspar and quartz which transect the schistose structure in the fine-grained groundmass. Most quartz crystals show strain shadows. The feldspar crystals, on the other hand, are not even fractured. Many contain clear, angular patches of feldspar surrounded by inclusion-filled feldspar with the same orientation, as though the clear patches were nuclei around which the crystal developed (Pl. 6, fig. 2). Some of the specimens contain leucoxene in small spots and patches resembling crystal outlines, and in quantities greater than in the normal porphyry. Contacts between foreign fragments and the porphyry can be observed in thin sections, but these are irregular, and feldspar crystals cut across them (Pl. 6, fig. 3). The foreign fragments are composed of a mat of small feldspar blades; though these feldspar blades do not appear to be oriented in sections cut normal to the schistosity, they are elongated and have their long directions more or less parallel in sections cut parallel to the schistosity. A few of the foreign fragments contain larger, isolated feldspar crystals similar to those in the surrounding porphyry. These are obviously unrelated to the bladed feldspar and developed after it.

Porphyry pebbles of the Timiskaming conglomerate.—The porphyry pebbles of the Timiskaming conglomerate likewise are composed of fresh porphyry similar to the normal porphyry of the region. The schistosity of the groundmass of this porphyry is marked chiefly by chlorite. The large crystals of quartz and feldspar are abundant and cut across the structure in the groundmass. The majority of feldspars are full of inclusions, and a few partially enclose groundmass material (Pl. 6, fig. 4). Some of the feldspars show good crystal outlines; others are amoeboid and composite under crossed nicols. A few feldspar crystals contain irregular fractures which have been rehealed by feldspar of the same composition. Scattered leucoxene spots are fairly abundant, and some of this material is included in feldspar. The contact of the pebble with its matrix is marked by streaks of chlorite. Feldspar crystals project across the contact and in some places have displaced the chlorite streaks (Pl. 6, fig. 5; Pl. 7, figs. 1, 2). An outstanding feature of the contact is its irregularity and lack of weathering or abrading; individual crystals are not cut by the contact.

Thus the Timiskaming conglomerate, like the Keewatin agglomerate, contains features which, similar to those in the main porphyry bodies, indicate the metasomatic development of quartz and feldspar. The striking resemblance between the porphyry of the agglomerate, the conglomerate and the stocks suggests a common source. Stratigraphic position, however, obviates derivation of agglomerate and conglomerate from the stocks if their age has been determined correctly.

Summary.—Lack of evidence of forceful intrusion of the porphyry bodies suggests that they may resemble igneous rocks but were actually formed by metasomatism. Supporting this is evidence that the quartz and feldspar crystals in the porphyry of the stocks, the ecole structures, the Keewatin agglomerate, and the Timiskaming conglomerate are metacrysts. Recapitulated, the evidence for the secondary development of quartz and feldspar is:

(1) Feldspar and quartz are not oriented uniformly with regard to any structure in the rock.

(2) On the contrary, many cut diversely across the schistosity (Pl. 4, fig. 6; Pl. 5, fig. 1).

(3) The crystals of feldspar and quartz show little strain and only slight evidence of fracturing.

(4) Both feldspar and quartz crystals contain abundant inclusions. In some, the inclusions are composed of leucoxene which preserves outlines of some pre-existing mineral (Pl. 4, fig. 1).

(5) Quartz crystals have tonguelike projections into the groundmass along the schistosity (Pl. 5, fig. 4).

(6) Locally, schist inclusions have been pushed aside by growing feldspar crystals (Pl. 6, fig. 6).

(7) Regrowth on rims of feldspar crystals is common. Fractures in crystals rehealed by feldspar of the same composition have been observed (Pl. 4, figs. 2, 3).

(8) Crushed and drawn-out feldspar crystals have been completely recrystallized (Pl. 4, figs. 4, 5).

(9) Embayed quartz crystals are common, but the embayments suggest encroach-

ment of the quartz on the groundmass (Pl. 5, fig. 2), not the opposite process characteristic of true igneous rocks.

(10) Feldspar crystals of the same composition as those in the porphyry stocks are found in the matrix surrounding the porphyry lenses of the ecole structures.

(11) Some feldspar crystals are composed of aggregates of smaller grains which extinguish and attain maximum brightness together under crossed nicols (Pl. 5, fig. 5).

Additional evidence of metasomatic origin lies in the lenses of carbonaceous porphyry which occur in the main porphyry bodies but lie on the strike of bands of carbonaceous schist.

INTERPRETATION[1]

Thus, the porphyry bodies have some characteristics of an intrusion but show many features suggesting replacement related to granitization. These two apparently conflicting observations are brought into agreement in the following interpretation.

After Keewatin folding an igneous rock intruded the steeply dipping beds of lava. The composition of the intrusive is a matter of conjecture, but it was presumably of an acid-alkalic nature because of the preponderance of sericite over chlorite in the present porphyry. The rock probably contained hornblende and sphene as both minerals were found in one thin section, and their outlines have been preserved in others by leucoxene. The evidence is too meager to speculate whether the intrusive was porphyritic. The Timiskaming folding has rendered this rock extremely schistose and has destroyed most of the original minerals.

Quartz and feldspar metacrysts were developed in the porphyry after the Timiskaming folding but before the formation of the quartz veins. The granite was intruded in the same period and may have given rise to solutions which aided in the formation of the metacrysts.

That hydrothermal solutions were instrumental in developing the metacrysts is suggested by the types of rocks in which they occur. Quartz and feldspar metacrysts are found not only in the stocks but also in the ecole structures, the matrix of the ecole lenses, the Keewatin agglomerate, and the Timiskaming conglomerate. It is significant that, aside from the stocks, metacrysts occur only in fragmental rocks. The Keewatin and Timiskaming slates and graywacke were apparently too fine-grained to have permitted the penetration of sufficient solutions to produce metacrysts. Moreover, the quartz and feldspar crystals in both the agglomerate and the conglomerate are fresh and abundant throughout. In contrast, in the porphyry stocks fresh, well-developed crystals occur only in certain sections; in other places feldspar crystals are either poorly developed or absent, or are sporadically distributed. This suggests that solutions deposited feldspar only in those sections which were more open to penetration. The direct factors controlling the development of feldspar, however, still remain obscure.

[1] At about the time the writer commenced his study certain features of the porphyry at the Dome mine attracted the attention of T. C. Holmes. It was some time after this that Doctor Holmes and the writer met, and both had formed essentially the same main conclusions regarding the origin of the porphyry. The reader is referred to Doctor Holmes' paper (1944).

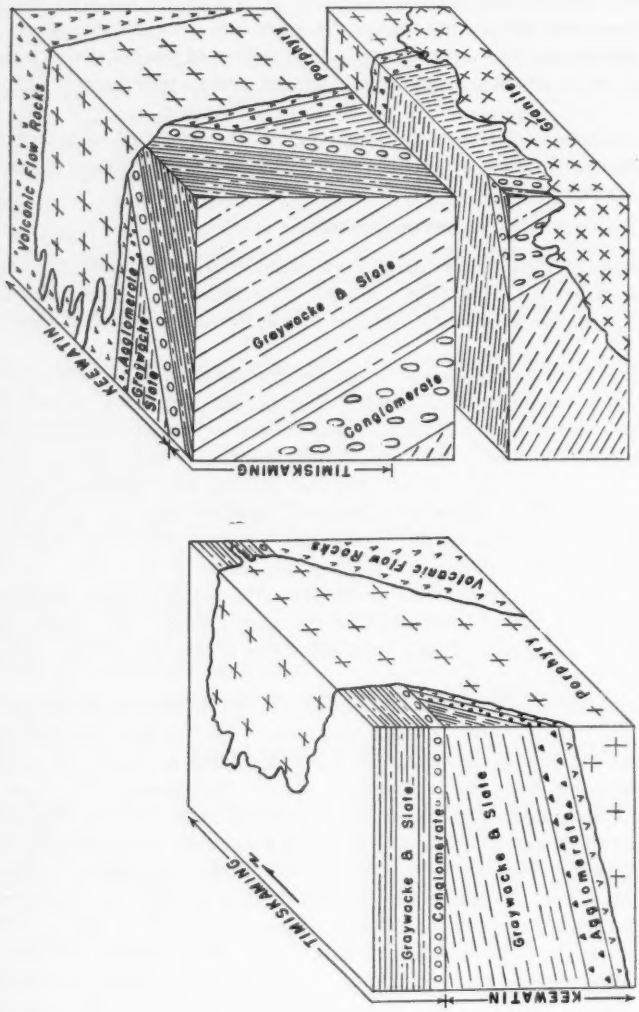

FIGURE 2.—*Diagram illustrating history of Porcupine rocks*

A, Diagram showing probable relations between the porphyry intrusive and the Keewatin and Timiskaming rocks prior to the Timiskaming folding. B, Diagram of the present relations between the porphyry and the Keewatin and Timiskaming rocks. Deep granite intrusion inferred as probable source of solutions which deposited quartz and feldspar crystals.

After the development of the quartz and feldspar crystals the porphyry rocks were fractured. This feature is well illustrated on the 4175 level of the McIntyre mine, where several intersecting fractures are sealed by tourmaline. The dips of the fractures are parallel to the linear schistosity, and the acute angles between them face east-west (Pl. 2, fig. 3). Corresponding fractures on a minute scale can be observed in some thin sections. These fractures contain sericite which is slightly coarser than that marking the schistosity. Still later, veinlets of quartz and carbonate cut the porphyry in some places. Carbonate has penetrated the rock beyond the veinlets and has partly replaced large areas of porphyry. This mineral is obviously very late because it has attacked feldspars and occurs as irregular patches in the more schistose porphyry.

SUMMARY AND CONCLUSIONS

The porphyry rocks of the Porcupine district occur as a group of stocks and dikes which conform to the general regional structures. The ends of the stocks interfinger with the schistosity of the enclosing rocks, but the sides of the stocks are fairly regular. In addition, the porphyry is schistose, but the degree of development of schistosity varies from place to place.

Two periods of folding have affected the rocks of the district: (1) Keewatin folding tilted the beds to steep angles and developed flow and fracture cleavage. This was followed by the removal of a minimum of 8000 feet of rock before the Timiskaming sediments were deposited. (2) Timiskaming folding produced a linear schistosity in the Keewatin volcanic rocks, the porphyry, and the Timiskaming conglomerate.

The form of the stocks indicates intrusion into well-defined pre-existing structures, and the schistosity shows that the porphyry was involved in folding. A contact between porphyry and sediments at the Dome mine suggests that the porphyry intruded rocks as young as Timiskaming. Therefore, the porphyry must have been intruded after Timiskaming deposition. Since it was involved in folding it must have been emplaced at least before the close of Timiskaming deformation. The intrusion probably occurred near the end of Timiskaming folding and may thus have been a forerunner of the larger and more widespread intrusions which were emplaced after the folding had ceased. Such a history for the region is shown diagrammatically in Figure 2.

Quartz and feldspar metacrysts developed hydrothermally after the Timiskaming folding. The chief evidence for this is the lack of relationship of the crystals to the structures in the groundmass of the porphyry. At the same time hydrothermal solutions deposited quartz and feldspar crystals in the Keewatin agglomerate and the Timiskaming conglomerate. A deep granite mass may have been the source of the solutions. Quartz, carbonate, and tourmaline veins cut the porphyry and also the Algoman granite. With these are associated some of the ore for which the region is famous.

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EXPLANATION OF PLATES 4–7

PLATE 4.—PHOTOMICROGRAPHS

Figure

1.—Leuxoxene (black) preserving outline of pre-existing mineral. (Plain light, × 40).

2.—Fractures (white) in feldspar, filled with feldspar of the same composition and orientation as the host. (Crossed nicols, × 40).

3.—Feldspar crystal with clear rim and serrated border. (Crossed nicols, × 60).

4.—"Drawn out" feldspar in schistose porphyry. (Crossed nicols, × 47).

5.—Same field as Figure 4, with feldspar at extinction position. Note how feldspar blends into groundmass. Arrows indicate long direction of feldspar. (Crossed nicols, × 47).

6.—Feldspar crystal transecting schistosity which trends from upper right to lower left. (Crossed nicols, × 16).

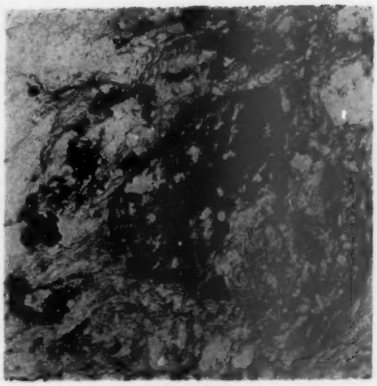

FIGURE 1

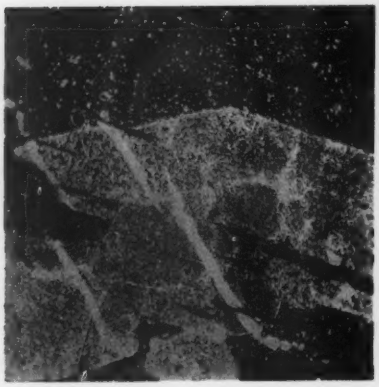

FIGURE 2

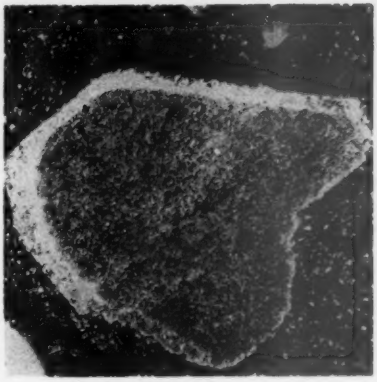

FIGURE 3

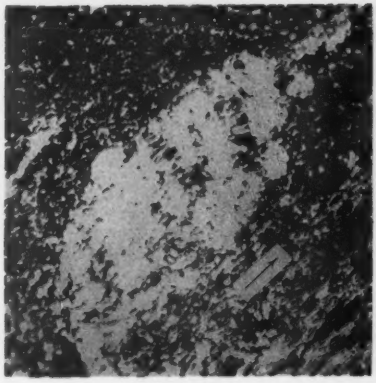

FIGURE 4

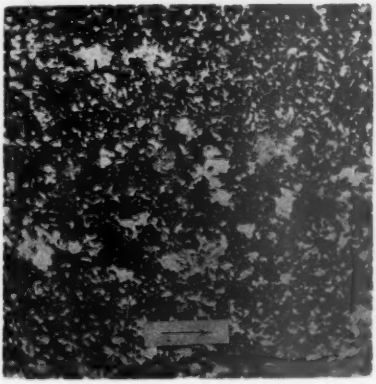

FIGURE 5

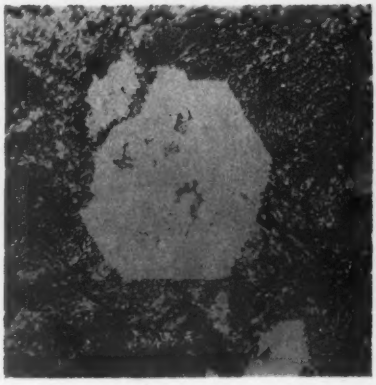

FIGURE 6

PHOTOMICROGRAPHS

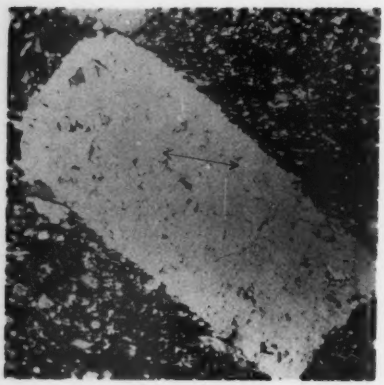

FIGURE 1

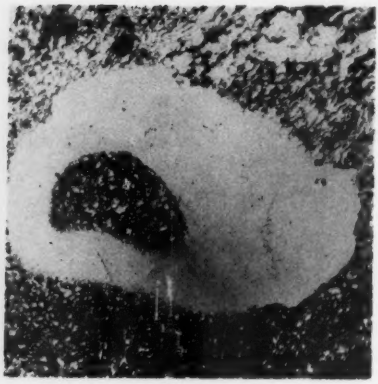

FIGURE 2

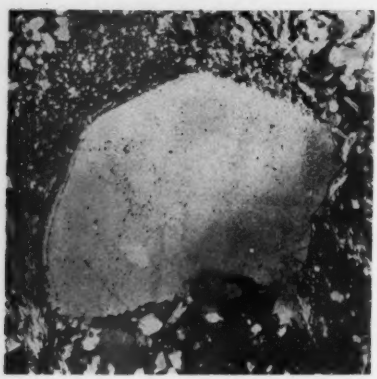

FIGURE 3

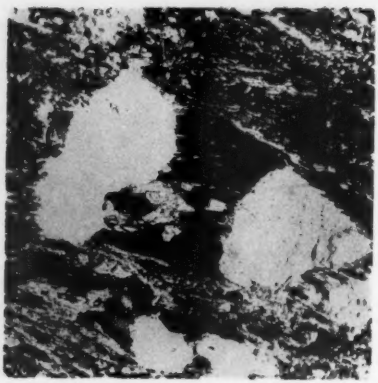

FIGURE 4

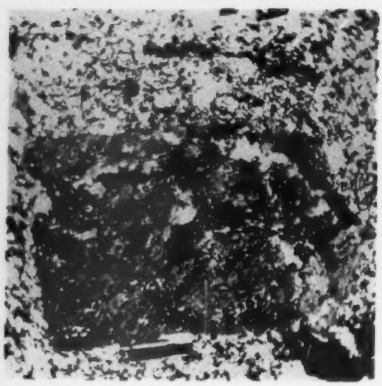

FIGURE 5

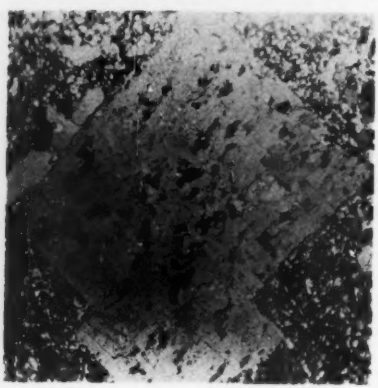

FIGURE 6

PHOTOMICROGRAPHS

PLATE 5.—PHOTOMICROGRAPHS

Figure

1.—Small feldspar tongues projecting from the edge of feldspar crystal into the groundmass schistosity. Note that the crystal lies at an angle to the schistosity, the direction of which is indicated by arrow. (Crossed nicols, × 40).

2.—Quartz crystal with included groundmass material. Note irregular margin. (Crossed nicols, × 57).

3.—Quartz filled with small inclusions. Note rim of inclusions on left side of crystal. (Crossed nicols, × 54).

4.—Quartz crystals with edges tonguing into the groundmass schistosity. (Crossed nicols, × 40).

5.—Composite feldspar crystal in porphyry. (Crossed nicols, × 42).

6.—Chlorite (black) enclosed in feldspar. (Crossed nicols, × 42).

PLATE 6.—PHOTOMICROGRAPHS

Figure

1.—Feldspar crystal in the chloritic contact zone between porphyry lens and matrix of ecole structure. The matrix, containing abundant leucoxene (black), shows in upper part of photograph. (Plain light, × 50).

2.—Irregular patches of clear feldspar (dark gray) in inclusion-filled feldspar. (Crossed nicols, × 45).

3.—Contact between porphyry of agglomerate and foreign fragment. Foreign fragment (top) contains more leucoxene (black spots) than porphyry. Note how feldspar (center) projects into foreign material and the accumulation of leucoxene at the feldspar's upper border. (Plain light, × 18).

4.—Feldspar crystal partly enclosing groundmass material. (Crossed nicols, × 45).

5.—Contact of porphyry pebble from Timiskaming conglomerate with its matrix (top). Large white crystal is carbonate, but smaller crystal to the right is feldspar. Both crystals project into fine-grained matrix material. Porphyry groundmass in lower part of picture is slightly coarser than material forming pebble matrix. (Crossed nicols, × 45).

6.—Porphyry in contact with schist inclusion. Dark streaks in inclusion are leucoxene. Growth of feldspar crystal has pushed aside schist. (Plain light, × 45).

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

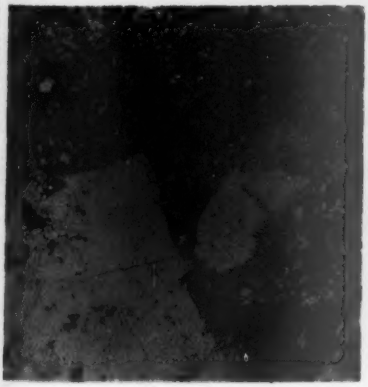

FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS

FIGURE 1

FIGURE 2

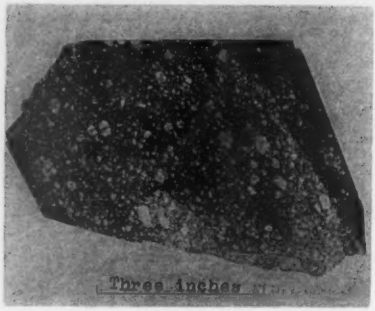

FIGURE 3

FIGURE 4

FIGURE 5

PHOTOMICROGRAPHS, POLISHED SURFACES

PLATE 7.—PHOTOMICROGRAPHS, POLISHED SURFACES

Figure

1.—Contact of a porphyry pebble (bottom) with its matrix (top). Dark-gray material is chlorite, and black is leucoxene. (Plain light, × 28).

2.—Same field as Figure 1 with nicols crossed. Feldspar crystal extends across contact. (× 28).

3.—Polished surface of porphyry lens from ecole structure. Feldspar crystals (white) occur in both the porphyry and the matrix material at right. Note the straight contact between porphyry and matrix, typical of exposures parallel to the schistosity.

4.—Polished surface normal to Figure 3. Contact between porphyry lens and matrix is irregular, and fragments in matrix show no distortion.

5.—Polished surface of matrix material on same specimen as above. Fragments show a directional trend from top to bottom, but feldspar crystals associated with them show no orientation.

(After Hurst, 1936. As the quartz prophyry intrusions are here regarded as part of the Algoman group, the columnar section has been altered somewhat from that originally published by Hurst, in which the quartz prophyry was included in the Keewatin.)